

Recent advances in polar low research: current knowledge, challenges and future perspectives

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ABSTRACT

Polar lows (PLs) are high-latitude intense maritime mesoscale weather systems that develop over open water near the sea ice margin or near snow-covered continents during cold air outbreaks. PLs pose a threat to coastal and island communities, transportation and offshore drilling platforms. PLs mainly develop during the cold season and their frequency exhibits a large interannual variability. Observations from polar-orbiting satellites are the main source of observational data to study PLs since conventional observations are sparse and unevenly distributed in high latitudes. PL forecasting has long remained a challenge due to the small size and short lifetime of these systems. Nevertheless, the representation of PLs in numerical models has significantly improved with the advent of high-resolution atmospheric models. Several studies have shown that baroclinic instability and convection play an important role in the development of PLs, but a thorough understanding of the physical mechanisms involved in the formation and intensification of PLs is yet to be developed. The relevant role of surface sensible heat flux and latent heat release in PL development has often been highlighted. The diabatic fluxes from the oceanic surface and associated with PLs can cause a decrease in the sea surface temperature (SST), whereas the strong wind speeds can lead to upper-ocean mixing in regions where an ocean temperature inversion is present. It is expected that global warming associated with anthropogenic climate changes may lead to an increase in the static stability of the atmosphere, thus affecting the climatology of PLs. In the North Atlantic the regions of PL activity will shift northwards as seasonal sea-ice margins migrate towards higher latitudes areas, and the frequency of PLs will decrease. Although our knowledge about PLs has significantly increased during the last decades, there are still many unanswered questions. Among the most pressing issues in PL research are the need to determine the objective criteria that define PLs and to devise an international intercomparison project of PL detection and tracking.

Keywords: polar low, severe weather, marine cold air outbreak, Arctic, Antarctic

1. Introduction

Polar lows (PLs) are high-latitude intense maritime mesoscale weather systems with lifetime ranging from three to 36 hours (Renfrew, 2003). They mainly develop over open water near the sea-ice edge and near the snow-covered continents during marine cold air outbreaks (MCAOs). The fact that MCAOs tend to be more intense in the Northern Hemisphere than in the Southern Hemisphere (Fletcher et al., 2016) probably explains why PLs are more intense in the Northern Hemisphere (Stoll et al., 2018). PLs are associated with severe weather such as

gale-force winds that can reach hurricane force ($\geq 33 \text{ m s}^{-1}$), and heavy snow showers or intense blizzard events. Consequently, they pose a threat to coastal and island communities, marine and air transportation, and offshore platforms in subarctic and Arctic regions. The harsh weather associated with PLs has long been endured by sailors and coastal populations living at high latitudes. Among other impacts, PLs have likely caused the loss of a large number of small vessels (Turner et al., 2003) and cases of PLs having caused the loss of human life have been documented. On 28 December 1986, following a derailment of a train on the Amarube Bridge in western Honshu, in Japan, due to strong winds associated with a PL, six persons deceased and six

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others were injured (Yanase et al., 2016). On 31 October 2001, the storm Torsvåg rapidly intensified over the island Vannøya, in Norway, causing the capsizing of a small boat and the subsequent death of one of the crew members (Norwegian Meteorological Institute, 2013). While PLs do not always cause such severe consequences, they often have an impact on the normal operations of the affected regions. PLs not only have important effects as high-impact meteorological phenomena, but also play a role as actors of the climate system; the strong surface sensible and latent heat fluxes from the ocean surface associated with PLs (e.g., Gachon et al., 2003) can in the long term act as forcing on the thermohaline circulation (Condrón and Renfrew, 2013).

For a long time, the origin of PLs was elusive. The incipience of polar-orbiting satellite imagery in the 1960s facilitated the advancement in the knowledge of PLs (Turner et al., 2003). Notwithstanding the large number of studies about PLs, a comprehensive and complete physical understanding of these systems is still lacking. The main challenges in the study of PLs are the lack of a universally accepted definition of PLs and the fact that atmospheric models and reanalyses do not always succeed at representing observed PLs. Consequently, their spatio-temporal distribution and the physical development mechanisms involved are not well understood yet.

This review article is motivated by the need to provide an updated review of the state of knowledge of PLs since the last comprehensive review, the book ‘Polar Lows: Mesoscale Weather Systems in the Polar Regions’ edited by Rasmussen and Turner (2003), was published almost twenty years ago. Moreover, during the 13th Polar Low Workshop, in 2016, the scientific community agreed that a new review article on PLs was needed due to the advancement of the field since the last review (Spengler et al., 2017). The publication of this review article also fits well within the current research themes of the Polar Prediction Project (PPP, Jung et al., 2016). The PPP, which spans from 2013 to 2022, was devised in the framework of the World Weather Research Programme of the World Meteorological Organization with the aim of promoting international cooperation in the domain of weather forecasting in the polar regions. Several studies of PLs have been published in the context of the Year of Polar Prediction (YOPP), a key activity of the PPP that consisted in the organisation of an intensive period of observation, modelling, verification and other activities.

This review is organised in seven sections. The next section describes the main characteristics of PLs and their climatology. The third section provides an overview of the main observational data used in PL studies, including conventional and satellite observations. The fourth section describes the main advances and current challenges in the numerical representation of PLs. The fifth section

presents the different mechanisms of formation and intensification of PLs. The sixth section describes the impact of PLs on the ocean and how climate change will affect the climatology of PLs. The last section focuses on the challenges and future perspectives of PL research.

2. Characteristics of PLs

2.1. Definition

PLs are part of the wider family of polar mesoscale cyclones (PMCs), which are defined by the European Polar Low Working Group (EPLWG) as ‘meso- α and meso- β -scale cyclonic vortices poleward of the main polar front’ (Heinemann and Claud, 1997), where the meso- α and meso- β scales correspond to 200–2,000 km and 20–200 km, respectively, according to Orlanski (1975). PLs are defined by the EPLWG as maritime PMCs whose horizontal scale does not exceed 1,000 km and which are associated with near-surface wind speeds of more than 15 m s^{-1} (Heinemann and Claud, 1997). Another definition that has been widely used in the literature is that of Turner et al. (2003):

‘A polar low is a small, but fairly intense maritime cyclone that forms poleward of the main baroclinic zone (the polar front or other major baroclinic zone). The horizontal scale of the polar low is approximately between 200 and 1,000 kilometres and surface winds near or above gale force.’

A few authors have suggested a classification of PLs in different types based on the synoptic environment where they form and their development mechanisms. Businger and Reed (1989) classified PLs in short-wave/jet-streak type, Arctic front type, cold-low type, and a combination of these three. The cold-low type included the meteorological systems known as ‘Mediterranean hurricanes’ or ‘medicanes’, which resemble hurricanes. Medicanes share in common with some PLs a spiraliform cloud signature and a warm core (Romero and Emanuel, 2017). PLs are now considered as distinct from medicanes, in particular because PLs form over cold arctic and sub-arctic environments, which is not the case for Mediterranean hurricanes, although they share some common characteristics with them. Both of them are relatively small-scale cyclones, and they develop in an environment with a large temperature contrast between the ocean surface and the atmosphere. Surface heat fluxes from the ocean and latent heat release play an important role in the development of both medicanes and PLs. Nevertheless, whereas the formation of medicanes requires the presence of an upper cut-off low with a cold core and large vertical extent, as well as important temperature contrast between the ocean surface and the atmosphere, the mechanisms of

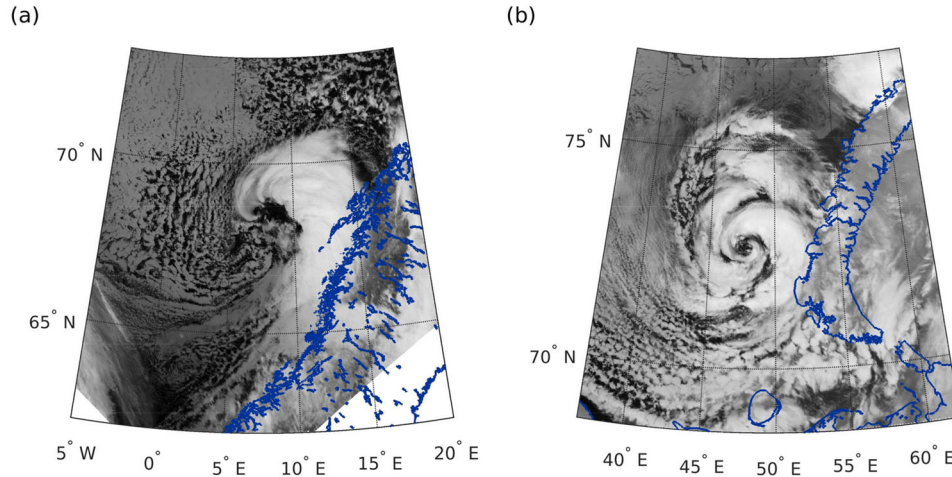


Fig. 1. IR satellite images of two PLs with different cloud signatures: (a) Comma-shaped PL over the Norwegian Sea on 25 March 2019 (image from AVHRR channel 4) and (b) Spiraliform PL over the Barents Sea on 20 December 2002 (image from MODIS spectral band 31).

PL genesis are more varied. Unlike PLs, medicanes peak season is autumn and early winter. With an average of two medicanes per year, these systems are much rarer than PLs (Romero and Emanuel, 2017). Rasmussen et al. (2003b) suggested a more detailed classification by dividing PLs into the following groups: reverse shear systems, trough systems, boundary-layer front, cold lows, comma clouds, baroclinic wave-forward shear and orographic PLs. Based on their cloud signature, PLs can be divided into spiraliform and comma-shaped (Fig. 1). However, the cloud signature of PLs can change from one type to another during their lifetime (e.g., Fu et al., 2004). Although in the past researchers thought spiraliform and comma-shaped clouds had clearly different genesis mechanisms, it has been shown that these differences are not clear-cut (Carleton and Carpenter, 1990). There are some differences between spiraliform and comma-shaped PLs though. Comma-shaped PLs have higher cloud tops than spiraliform PLs, and spiraliform PLs are characterised by much larger ice water content (IWC) and larger effective radius in the upper region of the cloud compared to comma-shaped PLs (Listowski et al., 2020).

2.2. Main characteristics

The definition of PLs by the EPLWG and by Turner et al. (2003) do not always suffice to classify a given cyclone as a PL. One of the main challenges to detect and track PLs in reanalyses and atmospheric models is to determine the objective characteristics that allow the identification of PLs. In effect, the number of detected PLs is rather sensitive to the choice of parameters (e.g., Yanase et al., 2016). The main criteria that have been used to

differentiate PLs from other type of cyclones are based on their size, intensity, lifetime and trajectory, as well as the static stability of the atmospheric region where they develop. Following Turner et al. (2003), many authors consider that the diameter of PLs ranges between 200 km and 1,000 km (e.g., Stoll et al., 2018). Other authors, however, consider that PLs have no minimum size (e.g., Blechschmidt, 2008), and some consider that their maximum diameter is 500 km (e.g., Wilhelmssen, 1985). The intensity criterion is often based on the maximum near-surface wind speed or the maximum vorticity at 850 hPa. Among the thresholds used, 13.9 m s^{-1} (e.g., Chen and von Storch, 2013) and 15 m s^{-1} (e.g., Yanase et al., 2016) were used for the near-surface wind speed, and $6 \times 10^{-5} \text{ s}^{-1}$ (Zappa et al., 2014; Michel et al., 2018) and $10 \times 10^{-5} \text{ s}^{-1}$ (Yanase et al., 2016) for the maximum vorticity at 850 hPa. Although the aforementioned definitions of PLs do not include a minimum lifetime, some researchers have considered a minimum lifetime of 6 hours (Zappa et al., 2014; Yanase et al., 2016), 9 hours (Landgren et al., 2019a), 12 hours (Michel et al., 2018) and even one day (Xia et al., 2012; Pezza et al., 2016) in their PL tracking algorithm. Pezza et al. (2016) also discarded cyclones whose lifetime was longer than 40 hours. Some authors consider that PLs move southwards (e.g., Zappa et al., 2014), but this criterion is too restrictive since in the North Pacific some PLs move zonally (Chen and von Storch, 2013), and in the Nordic Seas a significant number of PLs move westwards and even northwards (Rojo et al., 2015). Finally, given that a great number of PLs form in an atmosphere with low static stability, a criterion often used is the difference between the sea surface temperature (SST) and the temperature at 500 hPa (T_{500}). Most authors require that $\text{SST} - T_{500} > 43 \text{ K}$

to classify a cyclone as a PL (e.g., Xia et al., 2012); however, this threshold excludes a significant number of PLs (Stoll et al., 2018). In effect, the $SST - T_{500}$ criterion allows distinguishing between PLs and extratropical cyclones, but the line between these two types of cyclones is not clear enough (Yanase et al., 2016). In some cases, this criterion excludes PLs that develop in environments characterised by low stability in the lower troposphere but with $SST - T_{500}$ values that do not exceed the 43 K threshold (Radovan et al., 2019). Given the diversity of environments in which PLs develop, the suitability of the $SST - T_{500}$ criterion has recently been questioned on the basis that this criterion could favour classifying certain types of lows, such as those associated with significant convection, as PLs (Spengler et al., 2017). For PL forecasting, the Norwegian Meteorological Institute (MET Norway) uses a threshold of 40 K for the $SST - T_{500}$ criterion (Landgren et al., 2019a).

Many studies have investigated the structure of PLs. The vertical extension of PLs is rather variable, with some PLs being shallow and others having a large vertical extension. For example, Claud et al. (2009) analysed a PL of less than 2 km deep, whereas Føre et al. (2011) studied a PL whose circulation extended up to the tropopause. Using radar and lidar satellite cloud products, Listowski et al. (2020) analysed the clouds of 82 PLs that occurred between 2006 and 2017. They found an average cloud top altitude of 5.7 ± 1.4 km and a maximum cloud top of 9 km. The authors also found a small but statistically significant positive correlation between cloud top and diameter. As for the inner structure of PLs, a great number of case studies have noted the development of a warm core (e.g., Fu et al., 2004; Sergeev et al., 2018), which is caused by adiabatic warming in downdrafts (e.g., Yanase et al., 2004), by warm air seclusion (e.g., Føre and Nordeng, 2012) or by sensible heat fluxes and latent heat release (Gachon et al., 2003). When the warm core is due to adiabatic warming in downdrafts, a cloud-free eye is formed (e.g., Yanase et al., 2004). Regarding the cloud composition of PLs, Listowski et al. (2020) found that in the PL cloud band virtually all the condensed phase was ice, with only 0%–4% being supercooled liquid water, whereas in the eye the supercooled liquid water fraction increased from 20% at roughly 1 km below the cloud top to 60% near the bottom. They also found that the average IWC of PLs increased from 0.02 g m^{-3} at the cloud top to $0.2 - 0.5 \text{ g m}^{-3}$ further down.

Based on the current knowledge, the following characteristic scales of PLs emerge. Their horizontal scale is $L = 100$ km and their vertical scale is $H = 1$ km. Their horizontal translation speed scale is $U = 10 \text{ m s}^{-1}$, with most PLs between 4 and 13 m s^{-1} (Rojo et al., 2015; Smirnova et al., 2015). Thus, the advective time scale of PLs is $L/U \approx 3$ h. Based on these characteristic scales of PLs, we can

determine whether the hydrostatic and geostrophic approximations are appropriate for their study. The validity of the hydrostatic approximation in the study of a meteorological phenomenon depends on the aspect ratio of the system. If $H/L \ll 1$, the acceleration of the vertical speed in pressure coordinates is negligible compared with the vertical gradient of the pressure perturbation (Markowski and Richardson, 2010). For PLs, $H/L = 10^{-2}$, so the hydrostatic approximation is valid when studying the system as a whole. The validity of the geostrophic approximation is determined by the Rossby number $Ro = U/Lf$, where f is the Coriolis parameter. At the latitudes where PLs develop $f \sim 10^{-4}$, so $Ro = 1$, and the geostrophic approximation is not valid for the study of PLs.

2.3. Impacts on oceanic waves

PLs are associated with severe weather such as gale-force winds and heavy snow showers. Despite the relatively small horizontal extent of PLs, their strong winds can lead to the development of high waves due to the effect of the moving fetch on wave growth. If the wave group velocity is similar to the translation speed of the PL, the wave will be within the moving fetch for a long time. This phenomenon is known as ‘group velocity quasi-resonance’ or ‘extended fetch’ (Orimolade et al., 2016). Based on conventional observations of 29 PLs that passed over marine stations in the Norwegian and the North Seas, Rojo et al. (2019) found that the maximum significant wave height ranged from 3.2 to 11 m and, on average, the maximum significant wave height was 6.3 m. Orimolade et al. (2016) used a one-dimensional parametric wave model to determine significant wave height associated with 155 PLs. They found that extended wave development was possible in more than 50% (90%) of the PLs if they had an average translation speed of 10 m s^{-1} (8 m s^{-1}). For PLs moving at 8 m s^{-1} , the estimated maximum significant wave height was 9.13 m. There has been recorded at least one case of a rogue wave, which is an extreme storm wave, associated with a PL: the Draupner wave that occurred on 1 January 1995 in the North Sea. At the time when the rogue wave developed, the significant wave height of the sea was nearly 12 m. The Draupner wave reached 25.6 m of height (Cavaleri et al., 2016). Multiple PLs seem to generate larger maximum wind speeds and maximum significant wave heights than single PLs, although more data are needed to confirm these extreme wave features (Rojo et al., 2019).

2.4. Climatology

Owing to important differences between the Arctic and Antarctic environments, PLs that develop in the

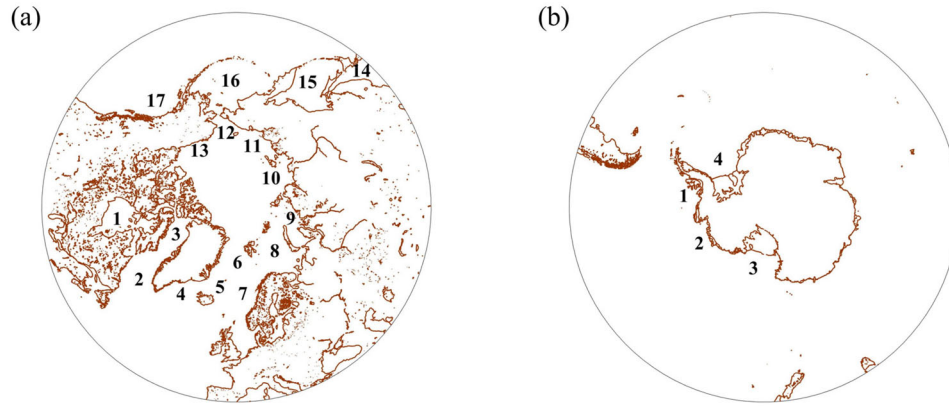


Fig. 2. Map of the marginal seas and other geographical features in the (a) Northern Hemisphere: (1) Hudson Bay, (2) Labrador Sea, (3) Baffin Bay, (4) Irminger Sea, (5) Iceland Sea, (6) Greenland Sea, (7) Norwegian Sea, (8) Barents Sea, (9) Kara Sea, (10) Laptev Sea, (11) East Siberian Sea, (12) Chukchi Sea, (13) Beaufort Sea, (14) Sea of Japan, (15) Sea of Okhotsk, (16) Bering Sea, (17) Gulf of Alaska, and (b) Southern Hemisphere: (1) Bellingshausen Sea, (2) Amundsen Sea, (3) Ross Sea, (4) Weddell Sea.

Northern Hemisphere and in the Southern Hemisphere possess distinctive characteristics, mainly due to major differences in the surface conditions and in the distribution of the continental and water masses (Fig. 2). Given that PLs develop during MCAOs, the climatology of PLs is closely related to their mean and variability features. The climatology of MCAOs has been studied using diverse indexes. The daily variability of the MCAO index depends almost uniquely on the lower troposphere air temperature, with no significant correlation between MCAOs and SST. In the northern North Atlantic and in the southern half of the Southern Hemisphere except for the Pacific Ocean, the seasonal variability of MCAOs is also primarily due to the variance of air temperature (Kolstad et al., 2009; Bracegirdle and Kolstad, 2010). In the southern Pacific Ocean, the seasonal variability of MCAOs is due to the variance of both air temperature and SST (Bracegirdle and Kolstad, 2010). MCAOs are more intense and more frequent in the Northern Hemisphere than in the Southern Hemisphere (Fletcher et al., 2016), mainly due to the presence of cold land masses in the winter months and high surface temperature gradients between the North American or Eurasian continents versus the adjacent seas during the cold season. Therefore, in the Antarctic the difference between the air temperature and the SST near the sea ice margin is smaller than in the Arctic and sub-arctic areas. Furthermore, the flow in the Antarctic is strongly zonal, even though at synoptic time scales the flow can become meridional (Rasmussen et al., 2003a). Although intense MCAOs are less frequent in the Southern Hemisphere, the most intense MCAOs in the Southern Hemisphere are as intense as their counterparts in the Northern hemisphere (Bracegirdle and Kolstad, 2010). Given that

MCAOs tend to be more intense in the Arctic, it is not surprising that overall the PLs that develop in the Southern Hemisphere are less intense than Northern Hemisphere ones (Stoll et al., 2018). Southern Hemisphere PLs are also smaller than the Northern Hemisphere ones (Turner et al., 2003).

Before giving a detailed description of the state of knowledge regarding the climatology of PLs, it is appropriate to give a short overview of the climatology studies of PMCs, which include PLs. The research community has developed a large number of climatologies of PMCs in the Northern Hemisphere (e.g. Harold et al., 1999; Michel et al., 2018) and in the Southern Hemisphere (e.g., Carleton and Carpenter, 1990; Verezhenskaya et al., 2017). For a long time, the research on PMCs was mainly focussed on the Arctic region since there was a lack of observational data covering a long time period in the Antarctic region (Rasmussen et al., 2003b). As a result, climatology studies of PMCs in the Southern Ocean have been developed relatively recently (Pezza et al., 2016). The identified climatologies have shown that PMCs seem to be ubiquitous phenomena in both hemispheres, whereas PLs are a small portion of PMCs. For instance, in the Southern Ocean, Verezhenskaya et al. (2017) found 1,735 PMCs during the austral winter of 2004, whereas Pezza et al. (2016) found 1,127 PLs between 2009 and 2012. In the Northeast Atlantic, two climatologies covering the same time period and a similar region found an average of 2,027 PMCs per year (Harold et al., 1999) and 69 PLs per year (Zahn and von Storch, 2008b). Michel et al. (2018) found just an average of 243 PMCs per winter in the Nordic Seas, but this low number may be explained by the strict lifetime criteria (≥ 12 h) applied to classify a cyclone as a PMC. Most of the PMCs in the Southern Hemisphere have a comma-

Table 1. Information about climatology studies of PLs developed using a subjective tracking method.

Author/s	Data used to identify PLs	Region	Period	Months
Wilhelmsen (1985)	Synoptic charts, conventional observations	Nordic Seas	1978–1982	All
Ese et al. (1988)	Synoptic charts	Norwegian and Barents Seas	1971–1983	All
Blechschmidt (2008)	IR imagery from AVHRR, wind speed from HOAPS	Nordic Seas	2004–2005	All
Noer et al. (2011)	MET-Norway PL list, IR imagery from AVHRR, wind speed from ASCAT and QuikSCAT, conventional observations	Nordic Seas	2000–2009	All
Rojo et al. (2015)	MET-Norway PL list, IR imagery from AVHRR	Nordic Seas	1999–2013	Oct.–Apr.
Smirnova et al. (2015)	Total water vapour and sea surface wind speed from SSM/I, IR imagery from AVHRR	Nordic Seas	1995–2009	Sep.–Apr.
Verezemskaya et al. (2017)	IR and water vapour satellite mosaics, wind speed from QuikSCAT	Southern Ocean	2004	June–Sept.

Only the data used to identify PLs is listed, even though some authors also used reanalyses to study the large-scale circulation associated with PL development.

shaped cloud signature (Carleton and Carpenter, 1990; Verezemskaya et al., 2017), which is at variance with the findings of Harold et al. (1999) that only a little over half of the PMCs in the Nordic Seas had a comma-shaped signature, the rest being spiraliform.

The climatologies of PLs provide an overview of the spatio-temporal distribution of PLs, even though there is still a lack of agreement regarding certain details such as PL frequency. These disagreements are due to a different selection of PL identification criteria, type of methodology used, and region and time period covered. Some researchers have used a subjective detection and tracking method to analyse the spatio-temporal distribution of PLs (Table 1), whereas others have used an objective method (Table 2). Bracegirdle and Gray (2008) have used a combination of a subjective and objective methods. Breakthroughs in computer science, especially in the domain of machine learning, have opened new possibilities for the detection and tracking of PLs. Recently Krinitskiy et al. (2018) have applied deep convolutional neural networks to detect the cloud signature of PMCs in the Southern Ocean, with promising results. The main challenges of using this method are the high computation cost and amount of time needed to train these neural networks and to optimise the hyperparameters.

Stoll et al. (2018) were the first to obtain a global climatology of PLs (Fig. 3). They applied a storm tracking algorithm to ERA-Interim (ERA-I, Dee et al., 2011) and the Arctic System Reanalysis (ASR, Bromwich et al., 2018), covering the periods 1979–2016 and 2000–2012, respectively, and compared their results to lists of PLs detected

with a subjective method. In both hemispheres, they found that PLs mainly develop during the cold season, this season being longer in the Southern Hemisphere (from April to October) than in the Northern Hemisphere (from November to March). Given the link between MCAOs and the development of PLs, these results seem coherent with the fact that between latitudes 55°N and 75°N, MCAOs are more frequent in winter and in autumn, and almost inexistent during summer, whereas between latitudes 55°S and 75°S the annual cycle of MCAOs is somewhat weak and the MCAOs can also occur in summer (Fletcher et al., 2016). The study of Stoll et al. (2018) also found that the Northern Hemisphere shows a maximum in the occurrence of PLs in January, when the temperature of the air masses over the North American and Eurasian continents reach the coldest values of the year (Serreze and Barry, 2014). In both hemispheres, the interannual variability of the PL activity is high. As for their location, PLs develop in the Northern Hemisphere down to latitudes of 40°N and 50°N in the North Pacific and the North Atlantic, respectively, and in the Southern Hemisphere up to latitudes of 50°S–60°S (Stoll et al., 2018). This geographical distribution is related with the main features of the seasonal sea-ice margins.

2.4.1. Northern hemisphere. PL activity is higher in the North Atlantic than in the North Pacific (Stoll et al., 2018; Fig. 3). In the North Atlantic, PLs form in the Nordic Seas¹ (e.g., Føre et al., 2011; Wu, 2021), the Denmark Strait (e.g., Kristjánsson et al., 2011b), the Labrador Sea (e.g., Moore and Vachon, 2002), the Davis

Table 2. Information about climatology studies of PLs developed using an objective tracking method applied to reanalysis data or simulated variables.

Author/s	Dataset/Model	Region	Period	Months
Zahn and von Storch (2008a)	COSMO-CLM	North Atlantic	1948–2006	All
Zahn and von Storch (2008b)	COSMO-CLM	North Atlantic	1993–1995	All
Chen and von Storch (2013)	COSMO-CLM	North Pacific	1948–2010	Oct.–Apr.
Yanase et al. (2016)	JRA-55	Sea of Japan	1979–2015	Oct.–Mar.
Pezza et al. (2016)	ERA-I AMPS	Southern Hemisphere	2009–2012	All
Stoll et al. (2018)	ERA-I ASRv2	Global Northern Hemisphere	1979–2016 2000–2012	All
Michel et al. (2018)	ERA-I	Nordic Seas	1979–2014	Oct.–Apr.

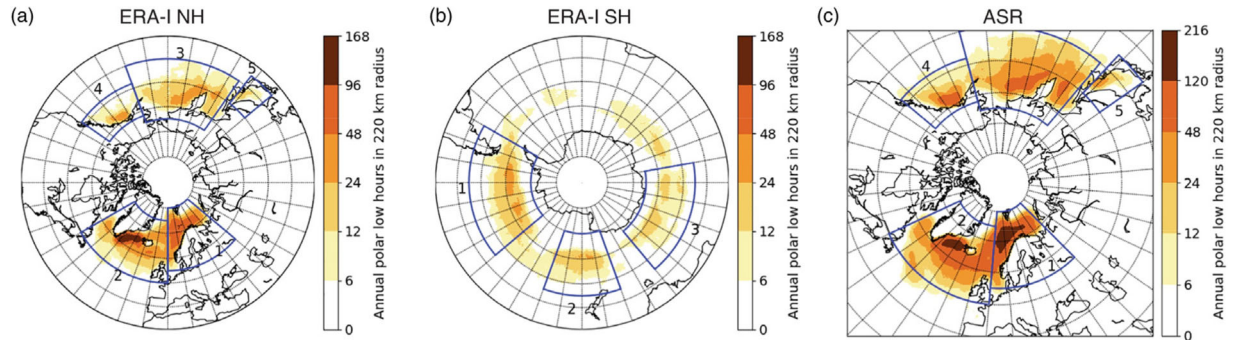


Fig. 3. Geographic distribution of PLs in the (a, c) Northern Hemisphere and (b) Southern Hemisphere, obtained applying an objective tracking algorithm to (a, b) ERA-I and (c) ASR reanalyses. The colour represents the annual average of PL duration within a radius of 220 km, which is calculated by multiplying the number of detected PL points by the temporal resolution of the reanalysis. From Stoll et al. (2018). © 2018 Royal Meteorological Society.

Strait (e.g., Roch et al., 1991) and the Hudson Bay (e.g., Gachon et al., 2003). In the North Pacific, PLs mainly form over the Sea of Okhotsk, the Sea of Japan (e.g., Fu et al., 2004), the Bering Sea and the Gulf of Alaska (e.g., Businger, 1987). PLs seldom develop over the marginal seas around the Arctic Ocean (Turner and Rasmussen, 2003); for instance, a PL has been observed in the Chukchi Sea (Inoue et al., 2010) and in the Kara Sea (Verezemskaya and Stepanenko, 2016) in the beginning of the PL season.

The best studied region is the Nordic Seas, where most PLs develop between September and May (Wilhelmsen, 1985; Ese et al., 1988; Blechschmidt, 2008; Bracegirdle and Gray, 2008; Zahn and von Storch, 2008a; Noer et al., 2011; Rojo et al., 2015). When looking at the monthly frequency of PLs, several authors have noted a minimum in February (e.g., Wilhelmsen, 1985), which seems to be associated with the presence of an anticyclone over Scandinavia (Noer et al., 2011). Nevertheless, some authors have found that the minimum of PL activity in February is not significant (e.g., Smirnova et al., 2015) or does not exist at all (Blechschmidt, 2008; Zahn and von Storch, 2008a). The annual frequency of PLs

varies greatly depending on the study. For instance, applying a subjective detection method, Smirnova et al. (2015) found an average of 45 PLs per year for the period between 1995 and 2009 in the Nordic Seas, whereas Noer et al. (2011) found an average of 12 PLs per year for the period between 2000 and 2009 in the Nordic Seas. What seems clear is that PL frequency exhibits a large interannual variability (e.g., Zahn and von Storch, 2010).

Studies of PLs in the Nordic Seas show that PLs develop near the sea ice margin (e.g., Harold et al., 1999) and in regions where SST is relatively high, particularly where the Norwegian Atlantic Current is found (Noer et al., 2011; Rojo et al., 2015). The regions where this current converges with the West Spitsbergen Current and the Barents Sea are particularly favourable for the formation of PLs (Rojo et al., 2015). In effect, these regions are characterised by a high MCAO index, which means that the atmospheric static stability is low (Kolstad et al., 2009). Thus, PL genesis occurs mainly between Iceland and the Finnmark county, in Norway, and over the Barents Sea (e.g., Wilhelmsen, 1985). PLs also form between the South of Greenland and Iceland (Blechschmidt, 2008) in the Irminger sea, which is

another region with a high MCAO index (Kolstad et al., 2009) where the presence of high elevated terrain from the Greenland ice-sheet strongly affects the low-level flow (Doyle and Shapiro, 1999). Synoptic conditions favourable for PL formation are the presence of northerly winds (e.g., Businger, 1985), north-westerly winds over the Greenland sea (Mallet et al., 2013) and north-easterly winds over the Barents Sea (Ese et al., 1988; Noer et al., 2011). Most PLs move southwards (Wilhelmsen, 1985; Blechschmidt, 2008; Rojo et al., 2015), but some move eastwards, westwards at even northwards (Rojo et al., 2015). Regarding the characteristics of PLs, different studies obtain somewhat differing results. For instance, the typical diameter and lifetime of PLs are respectively 300 km and 15 h according to Blechschmidt (2008), 350 km and 6–30 h according to Rojo et al. (2015), and 200–400 km and 9–18 h according to Smirnova and Golubkin (2017). The diameter of most PLs does not exceed 500 km (Smirnova et al., 2015), and few among them have a lifetime exceeding two days (Blechschmidt, 2008; Rojo et al., 2015). The steering level wind, which is the wind that dictates the translation direction of PLs, is located at 850 hPa (Rojo et al., 2015). With a characteristic propagation speed ranging from 4 to 13 m s⁻¹ (Rojo et al., 2015; Smirnova et al., 2015), most of them travel a distance of less than 1,000 km (Rojo et al., 2015). Regarding the weather conditions associated with PLs, Rojo et al. (2019) analysed conventional observations of 29 PLs developed over the Norwegian Sea and the North Sea between 1999 and 2013 to obtain their associated maximum wind speeds and associated maximum significant wave heights. They found that the maximum wind speed associated with the PLs ranged from 7 to 31 m s⁻¹ and, on average, the maximum wind speed was 17.1 m s⁻¹. They also found that, on average, large PLs are associated with higher maximum wind speed and higher maximum significant wave height than small PLs, and that fast-moving PLs are associated with higher maximum wind speed and higher maximum significant wave height than slow-moving PLs.

Some researchers have analysed the large-scale circulation patterns that are present during PL formation and intensification, in particular during wintertime weather patterns over the North Atlantic-Europe area (e.g., Mallet et al., 2013). In general, large-scale patterns leading to MCAOs are favourable for PL development. Blechschmidt et al. (2009) analysed the large-scale patterns associated with PL development in the Northeast Atlantic. They found that the large-scale circulation pattern associated with PL development over the Barents Sea is characterised by a low mean sea level pressure (MSLP) anomaly over the Barents Sea and a high MSLP anomaly over Iceland that tilts westwards with height. As

for the Norwegian, Iceland and Greenland Seas, the large-scale circulation pattern associated with PL development consists of a low MSLP anomaly over the Norwegian Sea and a ridge anomaly over Greenland and the Irminger Sea. The large-scale circulation pattern associated with the PLs that develop in the region between South Greenland and Iceland is characterised by a low MSLP anomaly east of Iceland that tilts westwards with height. These large-scale circulation patterns associated with PLs over the Nordic Seas and over the region between South Greenland and Iceland are similar, albeit somewhat shifted, to the respective patterns found during winter MCAOs in those regions (Kolstad et al., 2009).

Since PL development is related to specific large-scale circulation patterns, some studies have analysed the link between PLs and low-frequency variability patterns (Claud et al., 2007) or weather regimes (Mallet et al., 2013). The formation of PLs can be associated with North Atlantic weather regimes because the synoptic-scale anomalies that favour their development often have the same duration as those weather regimes. In the Norwegian Sea and the Barents Sea, PLs mainly develop during the Atlantic Ridge regime and the negative phase regime of the NAO (Mallet et al., 2013). This is in agreement with the small negative correlation between the annual wintertime MCAO index and the NAO index in the Barents Sea found by Kolstad et al. (2009). These authors did not find a significant correlation between the annual wintertime NAO index and the MCAO in the region comprising the Norwegian, Iceland and Greenland Seas. In the Labrador Sea, PLs primarily develop during the positive phase of the NAO and are virtually absent during its negative phase (Mallet et al., 2013), in agreement with the positive correlation between the annual wintertime MCAO index and the NAO index in the Labrador and the Irminger Seas (Kolstad et al., 2009). The positive phase of the NAO is not only associated with significantly low temperatures at 500 hPa, but also with northerly to westerly winds over the Labrador Sea and east of Canada, which are favourable for PL formation (Claud et al., 2007).

Another region that has received the attention of the scientific community is the North Pacific. Similarly to the Nordic Seas, most PLs develop during winter and their frequency is characterised by a high interannual variability (Chen and von Storch, 2013; Yanase et al., 2016). PLs are most frequent in the Western North Pacific. Specifically, they are most often found east of Japan, as well as in the Gulf of Alaska (Chen and von Storch, 2013). In the Japan Sea, PLs develop mainly west of Hokkaido, halfway between the Asian continent and Japan, and east of the North Korean coast. The PLs that develop over the Sea of Japan tend to move towards a

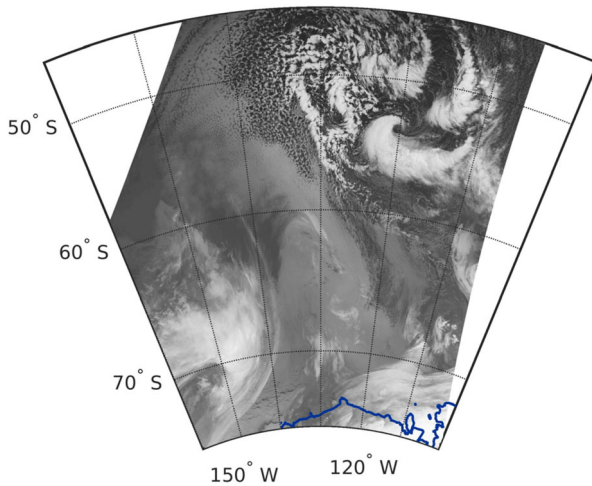


Fig. 4. AVHRR channel 4 image of a comma-shaped PL over the South Pacific Ocean on 14 July 2019.

direction somewhere in between the south and the east, and they reach their maximum intensity west of Japan. The steering level wind of the PLs that move southwards is located in the lower troposphere where north-westerly winds prevail, associated with synoptic cyclones and the winter Asian monsoon, whereas the steering level wind of eastward moving PLs is located in the mid-troposphere where mid-latitude westerly winds are present. The propagation speed of PLs that move eastwards is higher than that of the PLs that move southwards (Yanase et al., 2016). Regarding the frequency of PLs in the region, Chen and von Storch (2013) found an average of 172 PLs per cold season in the North Pacific and Yanase et al. (2016) found an average of 6.8 PLs per cold season over the Sea of Japan.

2.4.2. Southern hemisphere. In the Southern Hemisphere, PLs are more frequent during the winter season, although they are relatively frequent in autumn and spring, and a few develop in summer (Pezza et al., 2016). PLs mainly develop over the Bellingshausen Sea and along the eastern East Antarctica marginal ice zone (Verezemskaya et al., 2017; Stoll et al., 2018) as well as over the Amundsen Sea and the Southern Sea south of New Zealand (Stoll et al., 2018; Fig. 3). There is high PL occurrence along the Antarctic sea ice margin, which is likely due to the combined effect of katabatic winds and MCAOs (Verezemskaya et al., 2017). In effect, the mean MCAO index in winter has large values near the Antarctic sea ice edge (Bracegirdle and Kolstad, 2010). Although PLs are more frequent near the sea ice edge, they can also develop at lower latitudes (e.g. Fig. 4). Despite the fact that, on average, PMCs move south-eastwards, roughly 30% of PLs that develop in winter move

northwards. Explosive PLs are rare, constituting less than 1% of PLs (Pezza et al., 2016).

3. Observational data used in PL studies

3.1. Conventional observations and field campaigns

In the polar regions, surface observations are sparse and unevenly distributed in space. Most stations are concentrated in the more populated land regions, and they are mainly located along the coast (Casati et al., 2017). Therefore, the number of surface observations available for the study of PLs is limited. Surface observations from land meteorological stations have been used to study PLs affecting Norway (e.g., Hallerstig et al., 2021), Iceland (Kristjánsson et al., 2011b) and Japan (Fu et al., 2004), as well as more isolated regions such as the Hudson Bay (Albright et al., 1995; Gachon et al., 2003) and Kamchatka (Businger, 1987). Although upper-air stations are particularly sparse in high latitudes and the observation frequency is much lower than that of surface stations, these type of observations have also been used to study PLs (e.g., Kristjánsson et al., 2011b). Observations from maritime stations provide not only atmospheric variables such as MSLP, wind speed and temperature, but also SST and significant wave height associated with PLs (e.g., Rojo et al., 2019).

The most valuable source of PL observations are observational fields campaigns. Unfortunately, observational campaigns for the study of PLs using airborne instruments are rare, the first such campaign taking place in 1984 during the Norwegian Polar Low Project. Observational data of PLs have also been obtained by research flights during the Alaska Storms Program in 1987 and the Coordinated Eastern Arctic Experiment in 1989. The first time that low-level flights were performed in a region where a PL developed was during the Lofoten cyclone (LOFZY) field campaign in 2005 (Kristjánsson et al., 2011a). As part of the campaign, 21 drifting buoys, a research vessel and the research aircraft Falcon were deployed. The LOFZY campaign provided observations of two PLs that developed on 7 March 2005. The aircraft took measurements down to 30 m, which enabled the computation of surface sensible heat and moisture fluxes (Brümmer et al., 2009). A few years later, as part of the Norwegian International Polar Year (IPY) Observing System Research and Predictability Experiment (THORPEX, Kristjánsson et al., 2011a), a field campaign took place from 25 February to 17 March 2008. The main observational platform was the research aircraft Falcon that was equipped with probes for the *in situ* measurement of atmospheric turbulent fluctuations, a Doppler wind lidar, a Differential Absorption Lidar and

dropsondes. This campaign provided valuable observations of two PLs that developed on 3–4 March and on 16–17 March 2008, which were observed during three flights and one flight, respectively. The short-range forecast of the first PL was good, whereas the forecast of the second one was quite poor (Kristjánsson et al., 2011a). The PL that developed in the beginning of March is probably the most studied PL in the scientific literature (Linders and Saetra, 2010; Aspelien et al., 2011; Førre et al., 2011; Irvine et al., 2011; McInnes et al., 2011; Wagner et al., 2011; Førre and Nordeng, 2012; Stoll et al., 2020). The observations from the IPY-THORPEX campaign have also been used to evaluate simulations of a low-pressure system that developed during the IPY-THORPEX campaign but, despite the conditions being favourable for PL development, did not become a PL (Adakudlu and Barstad, 2011). More recently, the Aerosol–Cloud Coupling And Climate Interactions in the Arctic (ACCACIA) field campaign in March and April 2013 provided airborne and dropsonde observations of a PL developed on 26 March 2013 over the Norwegian Sea (Sergeev et al., 2017).

3.2. Remote sensing observations

Polar-orbiting satellite observations are the main source of observational data in high latitudes, particularly over the ocean. Products derived from the observations of passive instruments have proved very useful to study PLs. One of the satellite products that have been most widely used in PL research are Infra-Red (IR) satellite images, mostly IR calibrated radiances from the Advanced Very High Resolution Radiometer (AVHRR, e.g. Figs. 1a and 4) and from the Moderate Resolution Imaging Spectroradiometer (MODIS, e.g. Fig. 1b). Since the cloud signature of PLs can be identified in IR satellite images, these images have been often used to develop climatologies of PLs using a subjective tracking method (e.g., Rojo et al., 2015), to verify simulations (e.g., Sergeev et al., 2017) or simply to show the cloud signature of observed PLs (e.g., Furevik et al., 2015). More recently, Krinitskiy et al. (2018) have used IR satellite images as input for deep convolutional neural networks to detect PMCs. Products derived from the observations of microwave imagers and sounders have also been used in PL research. Microwaves penetrate through clouds and are only sensitive to clouds constituted by large ice particles; therefore, PLs associated with convective cells can be detected using satellite passive microwave radiometers (Claud et al., 2009). The Special Sensor Microwave Imager (SSM/I) products that have been used to study PLs are integrated water vapour (e.g., Smirnova et al., 2015) and surface wind speed (e.g., Martin and Moore, 2006). The Advanced Microwave Sounding Unit - A (AMSU-A), which is a microwave

temperature sounder, has been used to analyse the large-scale temperature field during PL development (Claud et al., 2009) and the warm core of a PL (Moore and Vonder Haar, 2003). The Advanced Microwave Sounding Unit - B (AMSU-B), which is a microwave humidity sounder, has been used to detect PLs whose development is mainly driven by convection since the scattering by large ice particles leads to low brightness temperature (Claud et al., 2009).

Products derived from the observations of active instruments have also been commonly used to study PLs. Scatterometers are undoubtedly one of the most valuable instruments to obtain information about the ocean surface wind field associated with PLs. Unfortunately, the current horizontal resolution of scatterometers is not high enough to capture strong wind gradients, and their temporal resolution is too low to capture the evolution of cyclones that evolve quickly (Bourassa et al., 2013). The main ocean surface wind products that have been used in PL research are those derived from the observations of SeaWinds scatterometer on board the QuikSCAT satellite (e.g., Kolstad and Bracegirdle, 2017) and the Advanced Scatterometer (ASCAT) (e.g., Furevik et al., 2015; see also Fig. 5). Ocean surface winds can also be derived from the observations of synthetic aperture radars (SARs), which have higher spatial resolution but smaller temporal resolution than scatterometer wind products. However, no SAR wind product is directly available (Furevik et al., 2015), at least at a global scale. As a result, SAR data has been used by a few researchers to obtain high-resolution data of the wind field associated with PLs (e.g., Martin and Moore, 2006) and to analyse the structure of convective cells (Hallerstig et al., 2021). The Cloud Profiling Radar, on board CloudSat, together with the Cloud-Aerosol Lidar with Orthogonal Polarisation, on board CALIPSO, have proven useful to obtain data about PL clouds such as the IWC, the liquid water content (LWC) and the cloud top height (e.g., Listowski et al., 2020). Forsythe and Haynes (2015) estimated that CloudSat/CALIPSO could capture an average of five PLs per year during the day.

Although less common, ground-based remote sensing is also a valuable source of PL observations. Ground-based weather radars provide reflectivity and reflectivity-inferred precipitation estimate data that allows the analysis of the precipitation field associated with PLs (e.g., Fu et al., 2004; see also Fig. 6). The infrasound signals detected in infrasound stations can be used to detect PLs since some of the processes associated with the development of the PL, such as convective turbulence, are among the possible sources an infrasound signal. However, for an infrasound signal to be detected, the environmental conditions need to be favourable for its propagation (Claud et al., 2017).

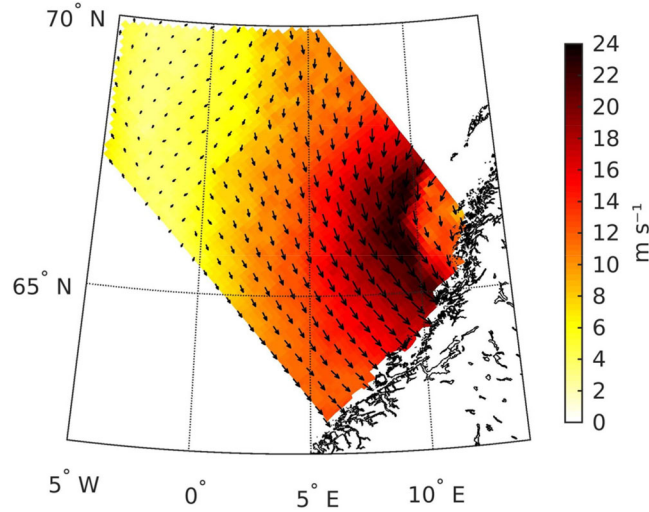


Fig. 5. Ocean surface wind speed (colour shading) and direction (arrows) derived from the observations of ASCAT on the 25 March 2019 at 18:40 UTC, when a PL (see Fig. 1a) was dissipating over the Norwegian coast.

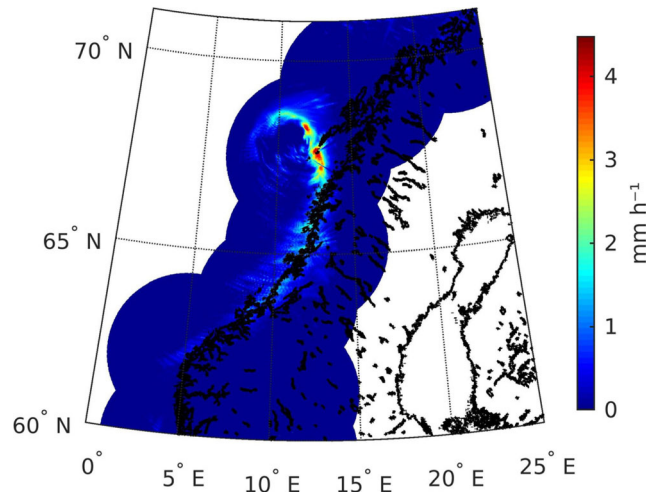


Fig. 6. Multi-radar mosaic of liquid water equivalent precipitation rate associated with a PL (see Fig. 1a) on 25 March 2019 at 15:00 UTC.

4. Numerical representation of PLs

In general, the performance of numerical weather prediction models over the polar regions is not very good due to, among other things, the lack of conventional observational data and an underperforming data assimilation. As a result, models do not correctly represent certain phenomena such as stable planetary boundary layers and thin clouds (Jung et al., 2016). For the simulation of PLs, compounding the challenge is the fact that PLs are small and short-lived systems.

The comparisons of simulation output against analyses, reanalyses and observations is a necessary step to evaluate the skill of the models at reproducing observed

PLs. Given that forecast verification against analyses is not adequate in the polar regions due to the strong influence that the model exerts on the analysis (Jung et al., 2016), satellite observations have been often used to verify the simulation of PLs. Unfortunately, the lack of high-resolution and high-quality observational data makes it difficult to evaluate convection-permitting models (Prein et al., 2017), which are the main atmospheric models used to simulate PLs. The type of method that has been the most used to verify PL simulations against observations is visual verification (e.g., Sergeev et al., 2017). To the best of our knowledge, Stoll et al. (2020) have been the first authors to use a spatial verification

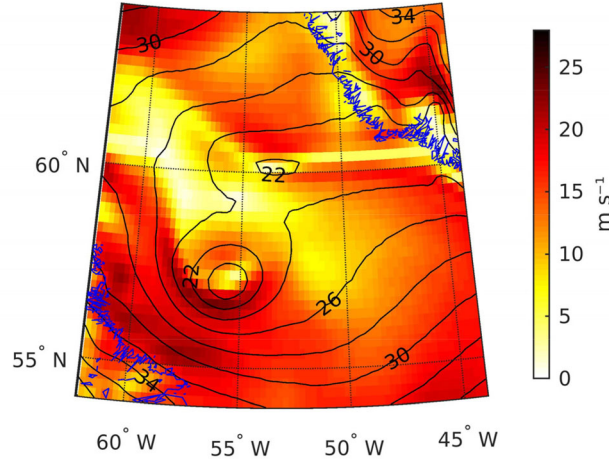


Fig. 7. Geopotential height (contour lines) and wind speed (colour shading) at 950 hPa on 8 March 2019 at 06:00 UTC from ERA5. The contour lines are drawn every 4 dam and the wind speed is in m s^{-1} . The geopotential height and wind fields show a PL over the Labrador Sea.

method to verify the simulations of an observed PL using a limited-area and a global atmospheric models. More specifically, they applied a ‘fuzzy’ verification method to verify their simulations against dropsonde observations. The results with the fuzzy verification technique showed that the limited-area model had better skill at simulating extreme values at small scales at the analysis time than the global model, whereas the standard verification statistics did not show this advantage of the limited-area model over the global one.

4.1. Reanalyses

The advances in global reanalyses, including the increase in resolution, has allowed a better representation of PLs. For instance, Laffineur et al. (2014) have shown that the ERA-I reanalysis with a grid mesh of 0.75° captures more PLs than the ERA-40 reanalysis (Uppala et al., 2005) with a grid mesh of 1.25° . Nevertheless, there are still some challenges in the representation of PLs in reanalyses, one of them being the correct representation of near-surface wind speeds. A large number of PLs are not captured by ERA-I since it underestimates near-surface wind speeds (Laffineur et al., 2014; Zappa et al., 2014); therefore, selecting a value of 15 m s^{-1} as threshold for the wind speed criterion, used to classify a cyclone detected in ERA-I as PL, excludes a significant number of PLs (Stoll et al., 2018). The underestimation of the near-surface wind by reanalyses seems to be a widespread issue. In their study of PMCs developed in the Southern Ocean, Verezhenskaya et al. (2017) found that, compared to QuikSCAT observations, the Climate Forecast System Reanalysis (CFSR, Saha et al., 2010), the Modern-Era

Retrospective analysis for Research and Applications (MERRA-2, Gelaro et al., 2017), the Japanese 55-year Reanalysis (JRA-55, Kobayashi et al., 2015) and ERA-I underestimate the maximum near-surface wind speed associated with PMCs. Whereas the underestimation by CFSR and MERRA is quite small, JRA-55 and ERA-I significantly underestimate the maximum wind speed, with differences of up to 10 m s^{-1} . A promising global reanalysis for the study of PLs is ERA5 (Hersbach et al., 2020; e.g. Fig. 7), which is the fifth-generation reanalyses of the European Centre for Medium-Range Weather Forecasts (ECMWF). ERA5 has a grid mesh of 31 km, a temporal resolution of one hour and so far covers the period starting from 1979, and recently released back to 1950.

Compared to global reanalyses, regional reanalyses seem to be more adequate to represent the phenomena characteristic of a particular region such as PLs. Moreover, since they tend to have higher resolution than global reanalyses, they are more likely to capture more small-sized PLs than global reanalyses. Smirnova and Golubkin (2017) found that the first version of the ASR, using a grid mesh of 30 km, captured more PLs than ERA-I, the reason being that in ASR the identification criteria regarding relative vorticity and surface wind speed were satisfied more often. They also found that the ability of ERA-I to capture PLs significantly increased with PL size, whereas the performance of the ASRv1 only decreased slightly for the smallest PLs. In contrast, Laffineur et al. (2014) did not find a relation between the PL size and the ability of ERA-I to capture PLs. PLs are probably better captured by the second version of ASR, which has a grid mesh of 15 km, since it is better than the

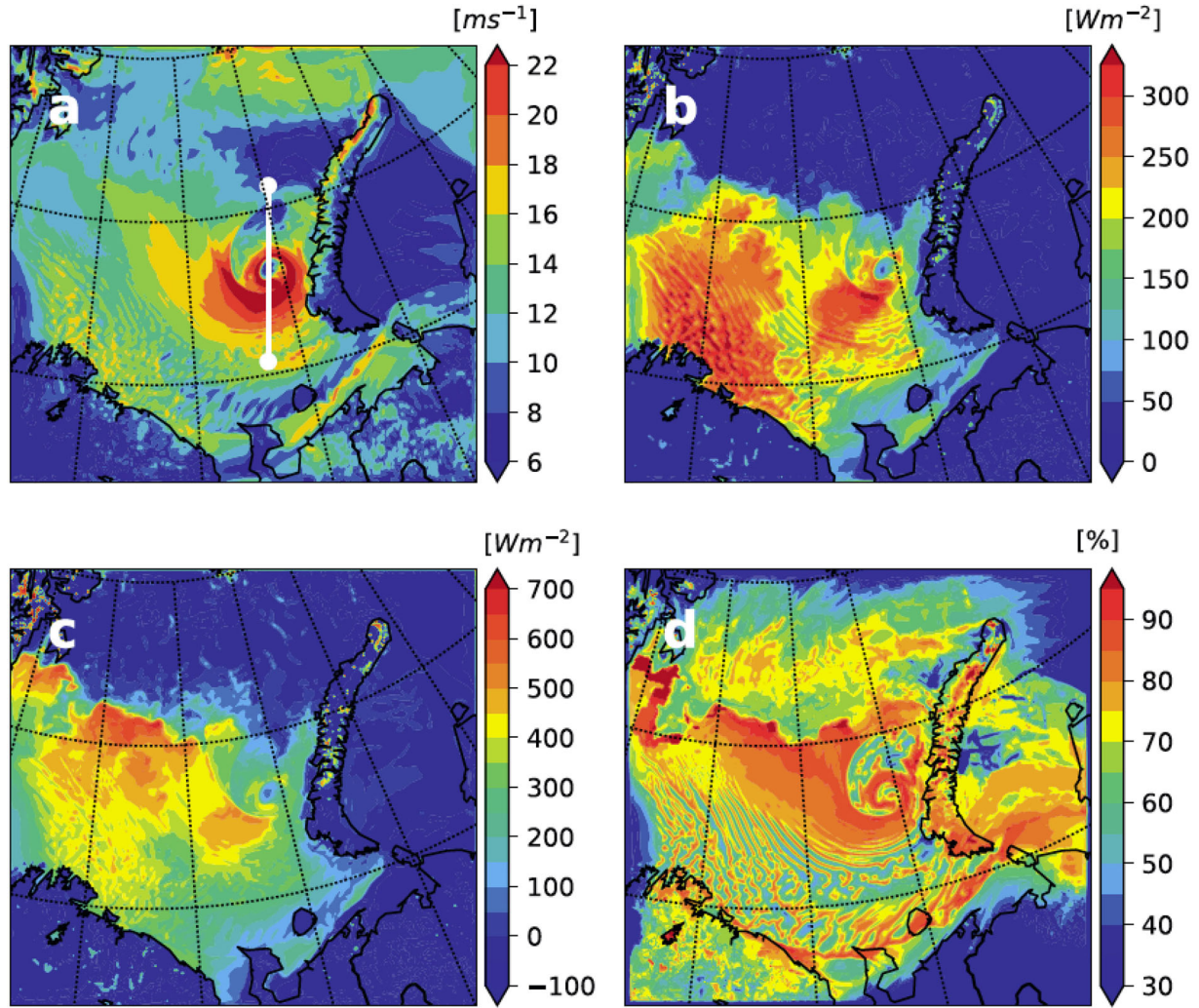


Fig. 8. PL developed over the Barents Sea (see Fig. 1b) simulated with the Weather Research and Forecasting (WRF) model: (a) mean wind speed at 10 m, (b) latent heat flux, (c) sensible heat flux and (d) relative humidity at 2 m on 19 December 2002 at 10:00 UTC (+34h). The SST and the sea ice concentration are provided by the NEMO model and are maintained constant during the simulation © 2020. Wu (2021). Published by Elsevier B.V. This is an open access article under the CC BY license (<https://creativecommons.org/licenses/by/4.0/>).

first version of ASR and ERA-I at reproducing near-surface atmospheric fields, including 10-m wind speed (Bromwich et al., 2018).

4.2. Atmospheric models

Until recently, global atmospheric models were too coarse to capture PLs and correctly represent their structure. For this reason, researchers have mainly used limited-area atmospheric models to study PLs (e.g. Fig. 8). A comparison of the performance of a global model and a limited-area model is provided by the study of Stoll et al. (2020), who simulated an observed PL with both the limited-area model AROME-Arctic, which has a grid mesh

of 2.5 km, and the global model ECMWF HRES, which has a grid mesh of 25 km. They found that both models showed good skill at simulating the PL, but the small-scale features, such as cloud patterns, and processes, such as convection, were better represented by AROME-Arctic than by ECMWF HRES. Hallerstig et al. (2021) have also compared PL forecasts produced by the model ECMWF HRES with grid spacings of 5 km, 9 km and 18 km, with those produced by the AROME-Arctic. They verified the forecasts of 10 m wind speeds against *in situ* and ASCAT observations. The authors found that the AROME-Arctic model performed better than the other models only when the verification was made against *in situ* wind observations. Given that the ASCAT product

that the authors used had a grid spacing of 25 km, the worse performance of the AROME-Arctic over the ocean may have been due to the lower resolution of the observational data. Hence, the representation of PLs can be improved by performing dynamical downscaling with a high-resolution model, which can be either used operationally or initialised with reanalyses or simulation output from an atmospheric model. For instance, Laffineur et al. (2014) found that, out of 29 observed PLs, more PLs were represented by simulations with the model Meso-NH initialised with ERA-I (22) or ERA-40 (17), than by ERA-I (13) and ERA-40 (6). Similarly, Pezza et al. (2016) found that the Antarctic Mesoscale Prediction System (AMPS), which employs the Polar WRF and has a grid mesh of 0.5° , represented 46% more PLs than ERA-I.

The representation of polar lows in numerical models has significantly improved with the advent of high-resolution non-hydrostatic atmospheric models. High-resolution models allow a better representation of surface or low-level processes such as the forcing of near-surface winds by topography (Jung et al., 2016) and convection (e.g., Sergeev et al., 2017, 2018). The study of McInnes et al. (2011) illustrates this added value from the increase in horizontal resolution of atmospheric models. The authors used the Unified Model of Met Office (MetUM) with the different horizontal grid meshes – 1 km, 4 km and 12 km – to simulate two PLs observed during the fieldwork campaign IPY-THORPEX. Convection was simulated explicitly in the 1-km model, whereas it was parameterised in the 12-km model. In the 4-km model, the authors used a modified convection scheme permitting explicit simulation of large-scale clouds. For the PL that was captured by the models, the increase in resolution lead to an improved simulation of the PL due to a better representation of convective precipitation. Nevertheless, other studies that also used the MetUM to simulate PLs did not find any significant improvement on the simulation of the wind, temperature and pressure fields when decreasing the grid mesh from 4 km to 2.2 km (Sergeev et al., 2018) and from 2.2 km to 0.5 km (Sergeev et al., 2017), although they did note that convective cells were better represented at higher resolution.

Increasing the resolution of atmospheric models entails the need to adapt the parameterisations to the higher resolution. In the case of the Arctic, the models already struggle to correctly represent the processes underlying the transformation of the air masses exchanged between the Arctic and mid-latitudes (Pithan et al., 2018). One of the parameterisations that needs to be optimised is that of surface fluxes. In the polar regions, the spatiotemporal scales over which fluxes vary are smaller than elsewhere (Bourassa et al., 2013). Given the relevant role played by

surface sensible and latent heat fluxes in PL development, the correct parameterisation of these fluxes is essential to model PLs. The main sources of error of the bulk flux algorithms are the systematic errors in the transfer coefficient, which is a function of the wind speed and surface layer static stability, and the precision of the measurements. The occurrence of strong winds or large differences between the SST and the overlying air temperature lead to disagreements between the different bulk parameterizations (Bourassa et al., 2013). Since PLs develop during MCAOs and are associated with strong winds, the use of different bulk parameterizations to simulate PLs could lead to different results.

A better representation of PLs requires not only improving atmospheric models, but also the use of quality input data. Many authors have noted the significant impact that initial conditions have on the representation of PLs (e.g., McInnes et al., 2011; Wagner et al., 2011). In the case of small-scale weather systems, initial conditions uncertainties are associated with convective and mesoscale instabilities (Zhang et al., 2019). In particular, the initial conditions of moisture at the mesoscale are critical for PL prediction (Spengler et al., 2016). The resolution of the model used may also have an impact on the forecast error growth. Stoll et al. (2020) found that the forecast error growth was higher for AROME-Arctic than for ECMWF HRES, and the authors argued that it was probably due to perturbations originating from convective processes. In the case of limited-area models, it is also important to use good-quality boundary conditions and to make sure that the synoptic conditions are well represented since the synoptic conditions preceding PL formation are crucial for their development (e.g., McInnes et al., 2011). Given that upper-level anomalies often play an important role in PL development, the use of the large-scale spectral nudging (SN) technique to enforce the synoptic-scale conditions with a regional climate model (RCM) allows a better representation of PL development (Zahn et al., 2008). For simulations in forecast mode, Zahn et al. (2008) showed that the SN technique improved the representation of PLs when the lead time was two weeks. For simulations in climate mode, Akperov et al. (2018) found that the climatology of cyclones in the Arctic was better represented by the RCM that used SN of the wind as opposed to those RCM that did not use it. The results could be further improved by using SN of temperature in addition to wind.

Since PLs can have an important impact on coastal populations, it is essential to improve the forecasts of PLs. Given the uncertainties in atmospheric models originated from diabatic processes, it is more adequate to use ensemble forecasts for severe weather forecasting in high latitudes (Spengler et al., 2016). One way to improve

forecasts in data-sparse regions is to use targeted observations, which can help reduce the initial condition errors (Aspelien et al., 2011; Irvine et al., 2011). However, it has been shown that targeted observations do not necessarily improve PL forecasts. For instance, when the additional observations provided by the IPY-THORPEX campaign were used in the data assimilation cycle, deterministic and probabilistic forecasts of the PL developed in the beginning of March were improved (Aspelien et al., 2011; Irvine et al., 2011), whereas this was not the case for the PL developed mid-March (Aspelien et al., 2011). Moreover, the improvement in the forecast may be modest compared to the spread between operational ensemble forecast members (Irvine et al., 2011).

It has been recently recognised that forecasting skill in the polar and sub-polar regions could be improved by using a coupled atmosphere-ocean-sea ice model (Jung et al., 2016; Spengler et al., 2016), in particular over the southern limit of seasonal sea ice margin in western Atlantic, i.e. the Gulf of St-Lawrence area (Pellerin et al., 2004; Smith et al., 2013). To the best of our knowledge, the only study that has used a coupled model to simulate a PL is that of Wu (2021). It showed that the simulation of the interactions between the atmosphere, ocean/ice and wave components allows a better representation of the surface heat fluxes, thus improving the representation of PLs during the stage of their development when the surface heat fluxes play a major role. Moreover, as might be expected, the high frequency interaction between these components determines the sea ice fraction and its dynamic (i.e. moving and growth at the sub-daily scale), which is another important variable involved in PL development.

5. Mechanisms of formation and intensification of PLs

5.1. Approaches for the study of PL development

The main hydrodynamic instability types of interest for the study of PLs are the baroclinic, barotropic and convective instabilities. As its name indicates, baroclinic instability develops in an environment where the air density depends on pressure and temperature. This type of instability is associated with the vertical shear of the average flow, which, as expressed by the thermal wind equation, is a function of the temperature horizontal gradient. Thus, baroclinic instability develops in regions characterised by strong horizontal temperature gradients and vertical shear of the horizontal wind, which is a common feature along the sea ice margin over polar and sub-polar areas in winter (Tansley and James, 1999). Barotropic instability develops in an environment where air density

only depends on pressure, which implies that isobaric surfaces are also isopycnic and isothermal. This type of instability is associated with the horizontal shear of the horizontal wind. Convective instability occurs in an atmosphere where entropy decreases with altitude, which corresponds to a statically unstable atmosphere.

For the study of the types of hydrodynamic instability that explain PL development, the approaches used are those typical of instability analysis, mainly the normal modes method and the initial value approach (van Delden et al., 2003). In addition, a large number of case studies of PLs, among them sensitivity studies, have been conducted with high-resolution atmospheric models to analyse the role of different atmospheric processes in PL development. The traditional sensitivity experiments consist of deactivating a specific process or processes throughout the whole simulation. Nevertheless, Yanase et al. (2004) have showed that deactivating a process throughout the whole simulation does not only have a direct impact on the development of the PL, but also has an indirect impact on the environment. Therefore, they have recommended that sensitivity experiments consist of deactivating a specific process during a short period of time at a particular moment during the simulation, thus ensuring that the environment will not be largely affected by the lack of that process. Following Yanase et al. (2004), Førø and Nordeng (2012) conducted sensitivity studies where they removed the physical process once the PL had developed with the aim of ensuring that the deactivation of the physical process would only have a direct impact on PL development. In contrast with the traditional approach of conducting sensitivity experiments on an observed PL, Adakudlu and Barstad (2011) conducted sensitivity experiments on a surface low that, despite developing in an environment favourable for PL development, did not become a PL. The aim of their study was to analyse whether they could trigger PL development by modifying the sea ice margin and the SST.

A useful approach to analyse the development mechanisms of PLs is the analysis of their energy budget. A few studies have been made on PL energetics, and the majority have used simplified atmospheric models and/or idealised initial and boundary conditions. For instance, Duncan (1977) conducted an energetics study of three PL cases with a two-dimensional linear atmospheric model. The author computed the eddy kinetic energy (EKE) and eddy available potential energy (EAPE) tendencies as well as the rate of conversion between different types of energy contributing to those tendencies. Other authors have used more complex atmospheric models, which are three-dimensional and non-hydrostatic, but with idealised atmospheric conditions such as cyclic boundary conditions (Yanase and Niino, 2007; Sergeev and Stepanenko,

2014; Terpstra et al., 2015). Given that PLs develop in a wide range of environments, the use of idealised conditions limits the representativeness of the results. Instead of using simulation output, Shimada et al. (2014) used operational analysis data from the Japan Meteorological Agency, whose grid mesh was approximately 11 km, to compute the EAPE and the EKE budgets of three PL with the aim of analysing the contribution of baroclinic and diabatic processes to PL development.

5.2. Polar low spectrum

The fact that PLs develop in different environments explains why different mechanisms can contribute to PL development. Baroclinic instability is a possible mechanism of PL development as PLs often develop in a shallow baroclinic atmospheric layer, mainly near the sea ice edge, and they can also develop in a deep baroclinic atmosphere (Terpstra et al., 2016). Barotropic instability is also a possible mechanism since some PLs form in an equivalent barotropic atmosphere (Businger, 1985). Convection can also contribute to PL development because the static stability of the atmosphere significantly decreases during MCAOs. A common synoptic situation leading to PL development in the Nordic Seas (Blechschmidt et al., 2009) and in the Sea of Japan (Yanase et al., 2016) is the presence of northerly winds on the west side of mid-latitude synoptic-scale cyclones, which produce MCAOs. Upper-level anomalies can also play an important role in PL development as shown by the fact that a cold trough or cold low at 500 hPa is often located over the region where PLs form in the Nordic Seas (Blechschmidt et al., 2009) and in the Sea of Japan (Yanase et al., 2016). In the Sea of Japan mature PLs moving southwards are located below an upper-level cold trough, thus favouring convection, whereas mature PLs moving eastwards are located in the east side of a upper-level cold trough, thus favouring baroclinic development (Yanase et al., 2016). Some researchers have tried to determine the large-scale environment associated with PL development in different regions. Such is the case of the studies conducted by Blechschmidt et al. (2009) and Bracegirdle and Gray (2008) who analysed the environments characteristic of PL development in the Nordic Seas, but reached partly contradicting conclusions. For PLs developing over the Barents Sea, Blechschmidt et al. (2009) found that the atmosphere was on average weakly baroclinic, whereas Bracegirdle and Gray (2008) found that the atmosphere was strongly baroclinic. These results emphasise the difficulty of trying to determine the characteristics of the environment leading to PL genesis in a particular region.

Between the end of 1970 and the beginning of the 1980s, there was an important discussion about the nature of PLs (van Delden et al., 2003). Whereas several authors affirmed that PLs were generated through baroclinic instability (e.g., Harrold and Browning, 1969), other stated that PLs were convective systems similar to tropical cyclones (e.g., Rasmussen, 1979). Several authors also suggested that baroclinic instability and convection can act together in the development of PLs (e.g., Sardie and Warner, 1985). Recent studies have confirmed that PL development is more complex than initially thought. Thanks to the valuable observations provided by the THORPEX campaign, it has been possible to study the PL that developed on 3–4 March 2008 in detail. Stoll et al. (2020) concluded that the development of this PL was divided into an initial baroclinic and a mature convective stage. Sensible heat fluxes and latent heat release played an important role in both stages: they contributed to the intensification of baroclinicity during the baroclinic stage, and they lead to a decrease in atmospheric stability during the convective stage.

As of today, there is still no widespread agreement on the importance of these different mechanisms in PL formation. Two conceptual models that have occasionally been used to explain PL development are the model developed by Montgomery and Farrell (1992), which was invoked by Grønås and Kvamstø (1995) to explain the development of two simulated PLs, and the Diabatic Rossby Vortex model, which has been invoked by Terpstra et al. (2015) to explain PL development in an ideal simulation with an Arctic moist-baroclinic environment and by Stoll et al. (2020) to explain how latent heat release affected the translation speed of an observed PL.

Given that several dynamic and thermodynamic factors play a certain role in PL development, and that the relevance of each factor varies depending on the case, it is convenient to refer to a PL spectrum in which PLs would be found with only a baroclinic or a convective origin, but also PLs formed as a result of the combined action of different forcing mechanisms (Rasmussen et al., 2003b). The idealised experience conducted by Yanase and Niino (2007) with a high-resolution non-hydrostatic model supports the idea of a continuum of PL formation mechanisms. The authors used simplified atmospheric conditions and imposed an axisymmetric vortex in gradient wind equilibrium as initial conditions. By applying different levels of baroclinicity for each experience, they noted that the structure and the dynamics of PLs varied gradually with the increase in the environment baroclinicity. In a barotropic environment, the development mechanisms of the vortex as well as its structure were similar to those of tropical cyclones. In an environment with strong baroclinicity, the meso- α -scale structure was similar to

that characteristic of a dry baroclinic instability wave. Moreover, the warming caused by condensation produced meso- β -scale structures showing a warm core near the PL centre.

5.3. Baroclinic and barotropic instability

Baroclinic instability mechanisms leading to PL formation appear in different forms. Several researchers have used the normal modes method to analyse the growth rate of perturbations of different wavelengths with the aim of finding the growth rate and wavelength corresponding to the PLs that develop fastest. Among the studies that have used the normal modes method, we find dry baroclinic models (e.g., Duncan, 1977; Reed and Duncan, 1987) and moist baroclinic models (e.g., Yanase and Niino, 2004). Duncan (1977) used a bidimensional quasi-geostrophic model to compute the normal modes of the unstable perturbations corresponding to three observed PLs. Among other approximations, this author neglected friction and the diabatic warming term. The results showed that the wavelengths of the unstable modes decreased as the static stability and the depth of the perturbation decreased. The model developed by Duncan (1977) was used by Reed and Duncan (1987) to analyse the serial development of four PLs. Noting that the intensification of the perturbations was due to a baroclinic mechanism, Duncan (1977) and Reed and Duncan (1987) concluded that the PLs that they had studied could be considered as shallow baroclinic waves. However, the model used was not capable of correctly reproducing all the characteristics of the observed PLs. The authors admitted that the quasi-geostrophic model, which neglects the horizontal variation of the static stability, was not adequate to analyse the development of PLs in the presence of important variations of the static stability. Reed and Duncan (1987) suggested that, after the initial baroclinic stage of PL development, deep convection likely played an important role in the development of the PLs. Using a non-geostrophic and non-hydrostatic model of a baroclinic flow, Yanase and Niino (2004) analysed the effect of convective warming on baroclinic instability. The authors concluded that the amount and spatial distribution of convective warming have a significant impact on the energy converted through the baroclinic mechanism.

All these studies have undoubtedly helped shape our understanding of PL development. Notwithstanding this, several models used were too simplified to enable a deep understanding of the development mechanisms of PLs. In general, the perturbations from which weather systems develop are complex and, subsequently, cannot be described as perturbations of only one normal mode

(Holton and Hakim, 2013). Idealised simulations can provide additional information on PL development. For instance, Terpstra et al. (2015) conducted idealised simulations in a moist baroclinic atmosphere, and they showed that, in the absence of an upper-level anomaly, friction, surface fluxes or radiation, a surface cyclonic warm-core perturbation can amplify in a moist baroclinic environment, with latent heat release being essential for the initial amplification of the perturbation. Simulations with three-dimensional atmospheric models are most suited for the analysis of the development mechanisms of PLs since all relevant processes are represented.

The baroclinic development of PLs can be triggered by forcing by shortwave upper-tropospheric troughs. A large number of case studies show that upper-level forcing by negative anomalies of geopotential height contribute to the development of PLs (e.g., Grønås and Kvamstø, 1995; Gachon et al., 2003). In some environments, upper-level forcing can be the main mechanism of PL formation. Mallet et al. (2013) concluded that during the weather regime known as ‘Scandinavian Blocking’, which is not associated with a significant anomaly of $SST - T_{500}$, an upper-level potential vorticity (PV) anomaly must be present to trigger PL development.

The PV approach has been often invoked to explain the development of PLs in situations where a upper-tropospheric PV anomaly is located over a low-level baroclinic zone (e.g., Føre et al., 2011; Shimada et al., 2014) or a near-surface PV anomaly (e.g., Moore et al., 1996; Verezhenskaya and Stepanenko, 2016), or both (Moore and Vachon, 2002; Martin and Moore, 2006). According to the PV framework, the vertical extension of the influence exerted by a PV anomaly is expressed by the Rossby depth, defined as $H_R = (fL)/N$, where f is the Coriolis parameter, L is the horizontal scale of the perturbation, and N is the Brunt-Väisälä frequency which represents static stability and is defined as $N^2 = g(d\ln\theta_0/dz)$, where g is the gravity acceleration and θ_0 is the potential temperature of the environment (Holton and Hakim, 2013). Accordingly, the vertical scale of a perturbation increases as the static stability decreases and as the horizontal scale of the anomaly increases. Therefore, for an upper air perturbation of a relatively small scale to produce baroclinic instability near the surface, static stability must significantly decrease and/or the vertical distance between the perturbation and the baroclinic zone must be relatively small (e.g., Shimada et al., 2014).

The PV approach can be used to explain the development of certain PLs. For instance, Moore et al. (1996) applied this approach to analyse a PL formed within a cloud band over the Labrador Sea. They explained that the generation of this PL was the result of an interaction between a high-level positive PV anomaly and a low-level

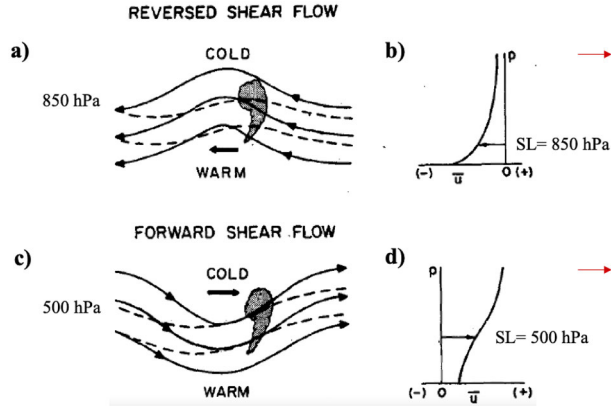


Fig. 9. Perturbation structure at the steering level in a (a) reverse-shear and (c) forward-shear flow. The solid lines represent the streamlines, and the dashed lines represent the isotherms. The thick black arrows represent the phase propagation vector and the steering level wind. The shaded grey region shows a comma-shaped cloud associated with the perturbation. The zonal component of the wind (u) at different pressure levels (p) is shown for the (b) reverse-shear and (d) forward-shear flow. The steering level wind is represented by the black arrows, and the direction of the thermal wind is schematically represented by the red arrows. Adapted from Businger and Reed (1989). © American Meteorological Society. Used with permission.

positive PV anomaly. Taking into account the fact that the horizontal scale of the high-level PV anomaly was roughly 500 km, and that the static stability over the Labrador Sea was very weak, the authors obtained a Rossby depth of around 8 km, which was sufficient to allow the interaction between both PV anomalies. The PV inversion method allows the quantification of the contribution of a certain PV anomaly to the temperature and wind fields. A particularly suitable method for the analysis of PL development is a piecewise PV inversion method, which consists of dividing the total PV anomaly field into three atmospheric layers to examine the contribution of discrete PV anomalies located in the upper troposphere, the lower troposphere and in a layer near the surface (Bracegirdle and Gray, 2009; Wu et al., 2011). The PV anomaly field can also be analysed at different scales to better understand the role that different scales of PV anomalies play in PL development (Shimada et al., 2014).

A type of PL that has attracted the attention of researchers is the reverse-shear PL (Fig. 9). This type of PL develops in regions where the steering level wind, which is normally located at the lower levels in the case of these type of PLs, is antiparallel to the thermal wind. In contrast to a typical baroclinic development, associated with a forward-shear flow, the perturbation tilts forward with height (forward meaning in the direction of

motion of the PL). In such systems, the warm air advection and the upwards movement take place upstream of the perturbation, and the cold air advection and the downwards movement take place downstream. Favourable conditions for the development of reverse shear PLs are found mainly in the Norwegian Sea, followed by the region south of the Denmark Strait, the Bering Sea and the Sea of Okhotsk (Kolstad, 2006). Accordingly, reverse-shear PLs have been commonly observed in the Nordic Seas (e.g., Blechschmidt, 2008), and they are much more frequent in the Norwegian Sea than in the Barents Sea (Terpstra et al., 2016). These PLs often develop over the maritime regions located west and north of Norway (Businger and Reed, 1989). In particular, one of the situations where these PLs are found is near the coast of Finnmark, where there are strong temperature gradients, when the wind blows from the north-east (Noer et al., 2011). In the Nordic Seas and in the Sea of Japan, reverse-shear PLs move southwards and forward-shear PLs move eastwards (Terpstra et al., 2016; Yanase et al., 2016). In the Nordic Seas, reverse-shear PLs move in a direction perpendicular to the SST isolines, towards regions of higher SST, whereas forward-shear PLs move following the SST isolines (Terpstra et al., 2016).

Contrary to baroclinic instability, the role of barotropic instability in PL development has received less attention. Only high-resolution models, with a grid mesh of a few kilometres, can correctly represent barotropic instability because the horizontal wind shear zone can be narrow (Leutwyler and Schär, 2019). Nagata (1993) showed that barotropic instability can significantly contribute to the formation of vortices at the meso- β scale. Even though barotropic instability has been observed in some PL case studies (e.g., Sergeev et al., 2017), it seems to play a secondary role (Businger and Reed, 1989).

5.4. Convection

The similarity between the structure of certain PLs and tropical cyclones (TCs) has led several authors to apply to PLs the theories developed to explain the intensification of TCs. The two main theories in the study of TCs are Convective Instability of the Second Kind (CISK) of Charney and Eliassen (1964), and Wind Induced Surface Heat Exchange (WISHE) of Emanuel (1986). The CISK and WISHE theories have several points in common, among them, the assumption of gradient wind balance and the use of an axisymmetric model. The two theories state that there is a positive feedback between the intensity of a cyclone and the convective warming by either the sensible and latent surface heat fluxes, according to WISHE, or the convergence due to friction in the

planetary boundary layer, according to CISK (Craig and Gray, 1996).

In order to develop the CISK theory, Charney and Eliassen (1964) assumed that the atmosphere is stable against large-scale convection, therefore in hydrostatic equilibrium, despite being unstable against small-scale convection. A hurricane is considered as a forced circulation that results from the latent heat released by the organised convection of the cumulus clouds. Taking into account these assumptions, the authors developed a quasi-hydrostatic two-level model in gradient wind balance and represented the flow as an axisymmetric vortex. The authors concluded that the amplification of an infinitesimal perturbation is the result of the combined effect of the convection associated with the cumuli and the TC. On the one hand, the TC provides humidity to the cumuli by means of the humidity convergence induced by the surface friction. On the other hand, by releasing latent heat associated with precipitation, the cumuli provide the TC with energy. This model has been applied to the study of PLs by, among others, Rasmussen (1979). Using a two-layer quasi-geostrophic and hydrostatic atmospheric model, the author found that the most unstable wavelengths computed with the model were in agreement with the size of the observed PLs, and that the intensification of the PLs was proportional to the contribution from the latent heat sources.

The CISK theory was criticised by Emanuel (1986) who highlighted that the tropical atmosphere is almost neutral to deep moist convection. In effect, the heat transport by convection quickly drains the convective available potential energy (CAPE) created, so that the CAPE could not play an important role in the maintenance of the cyclone. As in the case of the tropical regions, some authors affirm that the CAPE present in the polar atmosphere is constantly being drained due to the convection triggered by the strong sensible and latent heat fluxes from the ocean surface (Emanuel and Rotunno, 1989; Linders and Saetra, 2010). Accordingly, some authors have found low values of CAPE (Bracegirdle and Gray, 2009; Adakudlu and Barstad, 2011; Laffineur et al., 2014) and even no CAPE (Verezemskaya and Stepanenko, 2016) during the development of observed PLs. Van Delden et al. (2003) have argued that significant values of CAPE can occasionally be produced in polar regions as a result of the combined action of the surface fluxes and cold air advection in the higher layers of the atmosphere, and that CAPE does not disappear as fast as it is generated.

Emanuel (1986) developed an alternative theory to explain the intensification of TCs, which has been named WISHE. This author used a hydrostatic and axisymmetric model in gradient wind balance and imposed an initial

perturbation of finite amplitude. The author concluded that the maintenance of a TC in a stationary state is due to a positive feedback between the cyclone circulation and the sensible and latent heat fluxes from the ocean surface. By noting that certain environmental conditions favourable for the development of hurricanes are similar to those observed in the development of PLs, Emanuel and Rotunno (1989) decided to apply the WISHE model to the study of PLs. The results of their study showed a good agreement between the central pressure of the PL computed with the model and that observed in reality. Moreover, they noted that the PL structure as determined by the model was quite similar to that of the hurricanes, the main differences being that the PL showed a shallower circulation and more warming at the surface associated with the sensible heat fluxes from the ocean. In view of these results, they concluded that certain PLs can be considered as ‘Arctic hurricanes’, and that this type of PLs require a pre-existing perturbation whose amplitude must be equal to or larger than a critical amplitude.

To try to settle the debate regarding the importance of the CISK and WISHE mechanisms, Craig and Gray (1996) conducted several experiments with a non-hydrostatic and axisymmetric model with the aim of analysing separately the effect of modifying the moist and heat transfer coefficients, and the effect of modifying the quantity of momentum transfer coefficient. Their results showed that the intensification rate of PLs increased with the increase in the moist and heat transfer coefficients, whereas the modification of the momentum transfer coefficient had a small impact on the intensification rate, except when then modification of this latter coefficient was very large. Thus, the authors concluded that the WISHE mechanism explains PL intensification. Nevertheless, the CISK *versus* WISHE debate is not settled yet. In effect, some PL case studies have pointed at CISK as a possible mechanism involved in PL development (Sardie and Warner, 1985; Yanase et al., 2004), whilst others have pointed at WISHE (Albright et al., 1995; Bracegirdle and Gray, 2009; Føre et al., 2012).

The consideration of certain PLs as ‘Arctic hurricanes’ remains controversial. The intensity of PLs does increase with increasing SST (Kolstad and Bracegirdle, 2017; Stoll et al., 2020). Stoll et al. (2020) estimated that the maximum near-surface wind speed associated with an observed PL increased by 1 to 2 m s⁻¹ per K of increase in SST, and that this correlation seemed to be nonlinear for the highest SST. Nonetheless, sensitivity studies have shown that hurricane-like PLs can become as intense as hurricanes only when the SST is increased 6 K (Kolstad and Bracegirdle, 2017) and 8 K (Albright et al., 1995), which is unrealistic.

5.5. The role of surface heat fluxes and latent heat release

Sensible and latent surface heat fluxes can become very important during MCAOs, thus contributing to the development of PLs (e.g. Figs. 8b and 8c). For example, Føre and Nordeng (2012) found sensible and latent surface heat fluxes of $1,200 \text{ W m}^{-2}$ and 400 W m^{-2} , respectively, in the region of the eyewall of a simulated PL that reached a minimum pressure of 961 hPa. Many case studies have highlighted the importance of surface heat fluxes for PL intensification (e.g., Wagner et al., 2011), whereas a few case studies have found that surface heat fluxes played a negligible role in PL development (Sardie and Warner, 1985; Verezhenskaya and Stepanenko, 2016). Some case studies have highlighted that sensible heat fluxes played a more important role than latent heat fluxes in the intensification of a PL (e.g., Føre et al., 2012). Given that, according to the Clausius-Clapeyron equation, the saturation vapour pressure increases exponentially with temperature, it can be expected that the higher the latitude where PLs develop, the more important the role of sensible heat fluxes should be compared to latent heat fluxes and vice versa. This is confirmed by the study of PLs in the Sea of Japan conducted by Yanase et al. (2016). Regarding condensational heating, many case studies have found that latent heat release played an important role in the development of a PL (e.g., McInnes et al., 2011; Stoll et al., 2020).

Surface heat fluxes and latent heat release can not only directly contribute to PL development, but also indirectly by modifying, for example, the distribution of thermal and vorticity advection terms at low levels as well as the baroclinic structure along the sea ice margin during the mature stage of the PL (Gachon et al., 2003). In effect, several case studies have indicated that surface heat fluxes and latent heat release can contribute to the maintenance of the baroclinic environment necessary for PL development. Sensitivity studies conducted by Føre and Nordeng (2012) showed that condensational heating was crucial in supporting the initial baroclinic environment, whereas surface latent heat fluxes and, although less importantly, sensible heat fluxes contributed to the maintenance of the baroclinic environment during the mature stage of the PL. Surface sensible fluxes (Yanase et al., 2004; Stoll et al., 2020), latent heat fluxes (Yanase et al., 2004) and latent heat release (Wu et al., 2011) contribute to maintain the environment favourable for PL development by rendering the atmosphere less stable. As will be explained below, the surface heat fluxes also have an indirect impact on PL intensification through the modification of the SST (Wu, 2021).

5.6. The role of topography and sea ice

Topography is a factor that can facilitate the formation of PLs near, among other places, the Antarctic, Svalbard and Greenland. In the Antarctic, the katabatic winds play an important role in the generation of PMCs (Turner et al., 2003). In the case of Svalbard, Sergeev et al. (2018) found that the topography of the archipelago did not play a crucial role in the development of two observed PLs, although it provided a secondary source of positive vorticity on the lower level of the atmosphere that contributed to their intensification.

The role of Greenland's topography in the development of PLs has been noted since a long time (e.g., Businger, 1985; Sardie and Warner, 1985). The elevated terrain of the south of Greenland can alter the propagation of a synoptic cyclone, thus inducing a secondary circulation over the Labrador Sea that can contribute to PL development (Moore and Vachon, 2002; Martin and Moore, 2006). Another phenomenon that results from the topography of Greenland are katabatic winds, whose vertical extension does not normally exceed 300 m (Klein and Heinemann, 2002). The downhill direction of the katabatic winds is influenced by the Coriolis effect in regions where the topography does not impose too many restrictions to the flow. In the case of Greenland's coasts, the influence of the Coriolis effect is limited since the flow is canalised by the fjords and valleys (Serreze and Barry, 2014). This double effect of topography – the generation of katabatic winds and the imposition of restrictions to their flow – is evident in the study of Kristjánsson et al. (2011b), who conducted sensitivity experiments to study the impact of Greenland's topography on the development of a PL observed over the Denmark Strait. The PL did not form in the simulation where the whole Greenland surface was located at sea level, and therefore in absence of katabatic winds. In the simulation where only the high-level terrain of Greenland's east coast was removed, in addition to being weaker, the simulated PL developed more southerly than the observed PL, which could be explained by the influence of the Coriolis effect on the katabatic winds in the absence of restrictions to the flow.

In the case of Greenland's east coast, the intensification of the katabatic winds by a synoptic forcing can cause katabatic storms. Klein and Heinemann (2002) showed that these storms can play an important role in the formation of mesoscale cyclones. In a normal situation, the katabatic winds create favourable conditions for the formation of mesoscale cyclones. The flow convergence in the valleys causes the vertical stretching of vorticity, and the transport of very cold air towards the ocean contributes to the intensification of baroclinicity in the

lower layers of the atmosphere. The katabatic winds can be intensified by the forcing by a synoptic low positioned between Greenland and Iceland, resulting in katabatic storms. In effect, the winds associated with such a low cause an increase in the vertical extension of the katabatic winds, thus contributing to the vertical stretching of vorticity. Furthermore, there can be a positive feedback between the intensity of the katabatic winds and that of a mesoscale cyclone through the intensification of the pressure gradient between Greenland and the cyclone.

A factor that must be considered when analysing the effect of topography on the development of PLs is the adiabatic warming of cold air associated with katabatic winds. Whilst Klein and Heinemann (2002) have noted that the katabatic winds transport cold air towards the ocean, increasing low-level baroclinicity, Kristjánsson et al. (2011b) highlighted that, due to adiabatic warming of the cold air associated with katabatic winds, the temperature of this air becomes higher than the air temperature of the environment, thus contributing to the decrease in surface pressure.

Another factor that influences PL development is the sea ice cover. Sergeev et al. (2018) conducted a study to investigate the sensitivity of simulations of two PLs to the sea ice cover. They concluded that the sea ice cover near Svalbard exerts an influence on the intensification of PLs, but it is not a determinant factor of PL development. Similarly, Adakudlu and Barstad (2011) found that the modification of the sea ice edge around Svalbard did not trigger PL development of a low-pressure system that developed in the Barents Sea, despite the environmental conditions being favourable for PL development. The problem with these two experiments is that, given that the authors changed several features of the sea ice cover at the same time, it is not possible to attribute any changes in cyclone intensification to the sea ice extent, the orientation of the sea ice edge or its shape. In these experiments the modified sea ice edge was oriented zonally (Sergeev et al., 2018) or almost zonally (Adakudlu and Barstad, 2011), and was smoothed. Moreover, in both studies the SST was modified so that all the new ice-free ocean was assigned a SST of around 271 K, which was not realistic since the strong SST gradients remained in the original region instead of near the new sea ice edge. Some case studies indicate that the shape of the sea ice edge has an impact on PL development. An indented sea ice edge is more favourable for cyclogenesis than a straight sea ice edge (Albright et al., 1995; Gachon et al., 2003), because an indented sea ice margin generates a discontinuity in the sources or sinks of surface heat fluxes, resulting in a local Laplacian of diabatic and thermal forcing involved in low level trough or vorticity formation (Gachon et al., 2003).

6. PLs and climate change

6.1. Impact of PLs on the ocean

As atmospheric phenomena, PLs interact with the other components of the climate system. The impact of PLs on the ocean has long intrigued researchers. The surface heat fluxes associated with PL development directly cause a cooling of the ocean surface. The Greenland/Norwegian Seas and the Labrador Sea, which are regions of deep water formation, are frequently affected by PLs (Rasmussen et al., 2003a). By cooling the water of the ocean surface in these regions, PLs could modify the formation rate of the ocean deep water. For instance, intense storms in the Labrador Sea can cause major events of deep convection (Garcia-Quintana et al., 2019). Moore et al. (1996) were probably the first to highlight the possible existence of a coupling between PLs, whose time scale is short, and the thermohaline circulation, whose time scale is long. After computing the surface fluxes associated with a PL developed over the Labrador Sea, they concluded that the resulting water cooling was sufficient to trigger the sinking of water masses, thus contributing to deep water formation. Nevertheless, Isachsen et al. (2013) argued that the impact of PLs on the ocean circulation is rather weak. The authors computed the effective forcing period by PLs, defined as $T_f = \text{radius} / \text{speed of movement}$, and their results indicated that in the Nordic Seas the forcing of more than 50% of PLs over a portion of the ocean lasts less than six hours.

In spite of the small forcing exerted by individual PLs on the ocean, the whole forcing exerted by PLs and other PMCs that develop every year can have an important impact on the thermohaline circulation. Condrón and Renfrew (2013) analysed the impact of PMCs on the thermohaline circulation of the Northeast Atlantic with a global, eddy-permitting, coupled ocean-sea ice model. They conducted two simulations covering the period 1978 – 1998, one with parameterised PMCs and the other without them. Their results indicated that the increase in the surface heat fluxes associated with PMCs, with an associated decrease in the ocean surface temperature, caused the destabilisation of the water column. Consequently, the presence of PMCs caused an increase in the frequency and depth of ocean convection, as well as an increase in the area affected by convection. The authors concluded that the presence of PMCs induces an increase in the oceanic heat transport towards the Nordic Seas and a transport of deep water southwards through the Denmark Strait. A limitation of this study is that the authors used a coupled ocean-sea ice model, so the interactions between the ocean and the atmosphere were not represented. The recent study of Hirschi et al. (2020) has demonstrated that the atmosphere-ocean interactions

need to be resolved at high resolution in a coupled model as the Atlantic Meridional Overturning Circulation (AMOC) is quite sensitive to the model horizontal resolution. High resolution processes and models are needed to incorporate both ocean mesoscale eddies and atmospheric regional features, including PMCs and PLs, especially when high-frequency interactions between ocean/sea ice and atmosphere are key factors for the deep water formation as in the case of the AMOC (Garcia-Quintana et al., 2019).

The importance of the SST cooling caused by surface heat fluxes has been questioned by Sætra et al. (2008) who argued that the vertical mixing associated with strong winds may actually lead to an increase in SST. The authors showed that when strong ocean surface winds occur on a time scale of hours to a few days, the subsequent vertical mixing leads to a warming of the ocean surface in regions where the ocean surface water is colder than the water at depth. They highlighted that, although surface heat fluxes contribute to some cooling of the SST, their effect is weaker than that of vertical mixing. Nonetheless, their study has some limitations since they used a one-dimensional ocean model forced with constant values of surface wind speed, air temperature and relative humidity. Using a coupled atmosphere-wave-ocean/ice model to simulate a PL, Wu (2021) showed that PL development is indeed associated with both an increase and a decrease in SST. The surface heat fluxes that result from both the strong winds associated with the PL and the large difference between the SST and the overlying air temperature lead to a decrease in the SST. At the same time, in regions where there is an ocean temperature inversion (i.e., cold water near the surface and warm water in a deeper layer), such as near the ice edge, the upper-ocean mixing induced by the strong winds leads to the transport of warm water from a deeper layer to the sea surface.

6.2. Impact of climate change on the climatology of PLs

It is well known since a long time that the surface temperature shows higher variability, sensitivity and tendency in the Arctic region than at a global level. This phenomenon, known as ‘Arctic Amplification’, implies that global warming will be particularly intense and rapid in the Arctic (Serreze and Barry, 2011). Global warming will affect the sea ice and the snow cover as well as the mean and variability features of MCAOs. Since PLs form near the sea ice edge and near the snow-covered continents during MCAOs, we can expect that the spatio-temporal distribution and intensity of PLs will be affected by climate change.

6.2.1. Recent climate change. Over the last two decades, the Arctic surface air temperature has increased by more than twice the global average. Global warming has caused a decline in the Arctic sea ice extent for all months of the year, whereas the evolution of the Antarctic sea ice extent overall has not shown any statistically significant trend. The snow cover has also declined over continents bordering the Arctic (IPCC, 2019). Despite these changes, recent climate change seems to have had a relatively small impact on PL climatology. Michel et al. (2018) did not find any significant long-term trend in the frequency of PMCs in the Nordic Seas over the period 1979–2014. Similarly, Stoll et al. (2018) did not find any significant long-term trend in PL activity, neither in the Northern Hemisphere nor in the Southern Hemisphere, over the period 1979–2016. Nonetheless, they found a significant decrease in the activity of the most intense PLs in the Northern Hemisphere, which the authors attributed to a decrease in the intensity of MCAOs. There has been a small positive trend in PL occurrence in the North Pacific over the period 1948–2010 (Chen and von Storch, 2013). Given the few studies conducted on the impact of recent climate change on the climatology of PLs, it is not possible to draw any definite conclusion.

6.2.2. Future climate change. The reduction in the Arctic sea ice extent is projected to continue over the first half of the 21st century as a result of global warming (IPCC, 2019). According to most Coupled Model Intercomparison Project phase 6 (CMIP6) climate projections, the Arctic will become virtually sea-ice-free in summer before 2050 for the emission scenarios SSP1-1.9, SSP1-2.6, SSP2-4.5 and SSP5-8.5 (Notz and SIMIP Community, 2020). As for the Southern Hemisphere, there is low confidence in climate projections regarding the evolution of the Antarctic sea ice. Global warming will also continue to cause a decrease of snow on land in the Arctic (IPCC, 2019).

Global warming will lead to an increase in the static stability of the atmosphere in the North Atlantic (Zahn and von Storch, 2010; Mallet et al., 2017; Landgren et al., 2019b) and in the North Pacific, although the increase will be smaller in the latter (Landgren et al., 2019b). In the North Atlantic, the SST – T_{500} will decrease almost everywhere where PLs develop, and situations of high vertical instability will become less frequent (Mallet et al., 2017). Moreover, the loss of sea ice will lead to a change on the spatial distribution of atmospheric static stability. In effect, the static stability will decrease in some ocean regions that are presently ice-covered during winter and will become ice-free in the future (Mallet et al., 2017; Landgren et al., 2019b). According

Table 3. Information about the studies on the future climatology of PLs developed using dynamical downscaling and an objective tracking method.

Author/s	Driving data	Nested model	Region	Historical period	Future period	Scenarios
Zahn and von Storch (2010)	ECHAM5/MPI-OM	COSMO-CLM	North Atlantic	1960–1989	2070–2099	B1, A1B, A2
Landgren et al. (2019a)	CESM Large Ensemble	HCLIM-ALARO	Nordic Seas	1990–2005	2026–2035 2071–2080	RCP8.5

to Landgren et al. (2019b), in the Nordic Seas, excluding the Barents Seas, the decrease in the MCAO index will be stronger in December and January than in February and March, so the maximum MCAO index is projected to shift from January to February. In the Barents Sea, the opposite is projected to occur. In a future scenario with high greenhouse gas emissions, weather regimes in the North Atlantic are projected to have less influence on the local static stability of the atmosphere, represented by $SST - T_{500}$. Thus, the observed link between weather regimes and PL development may disappear in the future (Mallet et al., 2017).

The impact of global warming on the frequency and spatial distribution of PLs in the North Atlantic has been studied using dynamical downscaling (Zahn and von Storch, 2010; Landgren et al., 2019a) and statistical downscaling (Romero and Emanuel, 2017). Table 3 summarises the relevant information from the two studies of the future climatology of PLs developed using dynamical downscaling. Zahn and von Storch (2010) used a model with a grid mesh of 50 km, whereas Landgren et al. (2019a) used a model with a grid mesh of 12 km. Romero and Emanuel (2017; Fig. 10) used a complex statistical–deterministic method to analyse the evolution of the PL frequency between the periods 1986–2005 and 2081–2100 using the IPCC scenario RCP8.5 (for an overview of the representative concentration pathways, see van Vuuren et al., 2011). They generated a large number of synthetic PL tracks compatible with the historic and future climates simulated by CMIP5 models. All these studies found that in the North Atlantic the regions of PL activity will shift northwards as the sea ice extent decreases (Zahn and von Storch, 2010; Romero and Emanuel, 2017; Landgren et al., 2019a). The frequency of PLs will decrease by the end of the 21st century (Zahn and von Storch, 2010; Romero and Emanuel, 2017), and this decrease is observed for all months of the cold season, although there is high uncertainty since some climate projections show an increase in PL frequency (Romero and Emanuel, 2017). According to the best climate projections, the number of PLs will decrease by 15% on average (Romero and Emanuel, 2017). The decline in PL frequency will be due to the increase in the static stability

of the atmosphere (Zahn and von Storch, 2010). In the Nordic Seas, excluding the Barents Sea, PL frequency will decline in October, November, December and January, and will increase in March. There will be a reduction in the lifetime of PLs in December in the Barents Sea, and in November, December and January in the Nordic Seas (Landgren et al., 2019a). Regarding the future intensity of PLs, the studies do not seem to agree. Romero and Emanuel (2017) found that the probability of intense PLs will decrease (increase) in the west (east) region of the North Atlantic, whereas Landgren et al. (2019a) found a decrease in PL intensity for three winter months in the Nordic Seas.

Condrón and Renfrew (2013) hypothesised that the decrease in the frequency of PLs and their northward movement due to climate change would be followed by a decrease in deep convection in the Nordic Seas as well as a decrease in the formation rate of the deep water of the Greenland Sea and the reduction of the volume of the Denmark Strait Overflow Water (freshwater) that flows towards the North Atlantic. This would result in a weakening of the AMOC. In fact, the recent study of García-Quintana et al. (2019) points out the fact that a decrease in the storms crossing the Labrador Sea with a consequent reduction in the winter heat loss might be a bigger threat to deep convection and Labrador sea water formation in the future, than the expected increases in the freshwater input from the Arctic area.

7. Challenges and future perspectives

7.1. Detecting and tracking PLs

Although our knowledge about PLs has significantly increased during the last few decades, there are still many unanswered questions. During the last Polar Low Workshops organised by the EPLWG, the scientific community highlighted the need to determine the objective criteria that define PLs (Spengler et al., 2017; Heinemann et al., 2019). These objective criteria could be then used in detection and tracking algorithms to study the climatology of PLs. In this regard, during the 13th Polar Low Workshop the scientific community expressed the need to

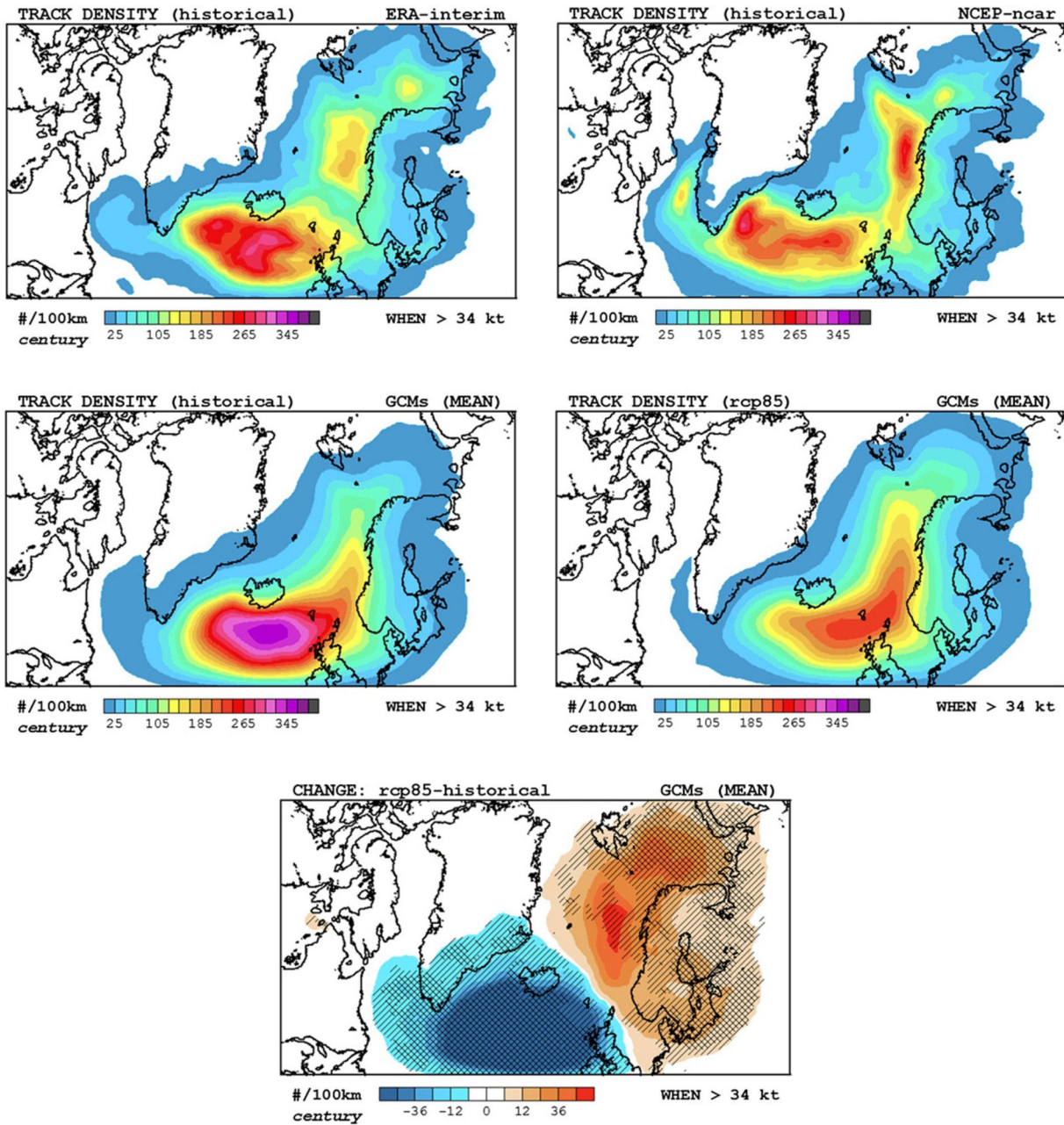


Fig. 10. PL track density obtained by applying a statistical downscaling method to (top left) ERA-I and (top right) NCEP-NCAR reanalyses, and to the multimodel mean of 20 Global Climate Models for the (middle row, left) historical and (middle row, right) future RCP8.5 scenarios. (Bottom) Change in PL track density between the historical and the future periods. Model agreement at 66% and 80% levels is indicated by the hatched and cross-hatched areas, respectively. The PL track density is the number of storms per century within a radius of 100 km. From Romero and Emanuel (2017). © American Meteorological Society. Used with permission.

create an international intercomparison project of PL detection and tracking (Spengler et al., 2017). PL research would enormously benefit from a project similar to the ‘Intercomparison of mid-latitude storm diagnostics’ (IMILAST) project, which is an intercomparison project of extratropical cyclone detection and tracking. Among

other achievements, the research conducted within this project has provided information about the characteristics of mid-latitude cyclones that are consistent across different tracking methods (Neu et al., 2013). Since some of the challenges for identifying and tracking mid-latitude storms, such as their different shapes and sizes, are

similar for PLs, a project similar to IMILAST applied to PLs could help unveil the characteristics of PLs that are consistent across different tracking methods. In the framework of the IMILAST project, Simmonds and Rudeva (2014) conducted an intercomparison of tracking methods applied to extreme Arctic cyclones, showing the potential of such an approach to analyse the characteristics of cyclones developed in high latitudes. In particular, there is a need to analyse the climatology of PLs in the Southern Hemisphere, which has received scarce attention so far.

7.2. *Observing and modelling PLs*

For a PL detection and tracking algorithm to successfully identify PLs, the data used as input needs to correctly capture PLs. Hence, the correct representation of PLs in analyses, reanalyses and in the output of atmospheric models constitutes another key challenge for PL research. In order to obtain a better representation of PLs, more observations and better models are needed. Given the sparseness of the conventional observations in the polar regions, improving the quantity and quality of observations cannot be achieved in the short term. However, observational campaigns focussed on PLs could provide data against which simulation output could be verified, which is a necessary step to improve high-resolution atmospheric models. In particular, more airborne campaigns to observe the three-dimensional structure of PLs are needed. Dropsondes observations can provide valuable data for the analysis of the development mechanisms of PLs (e.g., Føre et al., 2011) and for the verification of PL simulations (e.g., Stoll et al., 2020). Unfortunately, given the small size and short lifetime of PLs, the success of such observational campaigns strongly depends on the skill of the models at forecasting them. The need to improve PL forecasting is indeed another challenge highlighted during the last Polar Low Workshops organised by the EPLWG (Spengler et al., 2017; Heinemann et al., 2019). For weather in mid-latitudes, the lead time of a skilful deterministic forecast is roughly ten days. This practical atmospheric predictability limit could be improved by reducing the uncertainty on the initial conditions. A decrease in the uncertainty of an order of magnitude would result in an increase of up to five days in the lead time. Nevertheless, for small-scale weather systems the room for improvement is much smaller given that many initial conditions uncertainties are associated with convective and mesoscale instabilities (Zhang et al., 2019). Moreover, in contrast to forecasting in tropical regions and mid-latitudes, the sources of perturbation growth are not well known yet for the polar regions (Jung et al., 2016). Thus, an interesting course of research

would be to analyse the sources of perturbation growth in order to determine the atmospheric predictability limit for PLs. As far as atmospheric models are concerned, PL forecasting has benefitted from the development of high-resolution atmospheric models, although an increased resolution entails the need to adapt the parameterisations accordingly. In this regard, one of the current issues in operational forecasting is that the current size of the grid mesh of operational models, which is of a few kilometres, corresponds to the ‘grey zone’ of convection (Heinemann et al., 2019).

The use of high-resolution, limited-area coupled atmosphere-ocean-ice models to simulate PLs is a promising area for future research. The recent study by Wu (2021) has shown that a coupled atmosphere-wave-ocean-ice model provides an improved representation of PLs by accounting for the coupling processes that directly and indirectly affect PL development. Further, coupled models provide a more comprehensive picture of the impact of PLs on the ocean. In addition, coupled atmosphere-ocean-wave models are required to correctly represent how surface winds are modulated by ocean waves (Gutowski et al., 2020).

The upgrading of numerical models needs to be accompanied by appropriate methods to verify the simulation results against observations. In PL research, it has been customary to perform visual verification of PL simulations. Nevertheless, objective verification methods should be applied since they allow the quantification of model performance. Among the objective verification methods available, spatial verification methods are preferred over traditional verification statistics because they are more suitable for the verification of spatial forecasts. Feature-based verification methods are particularly suitable for the verification of PL simulations since they provide information on the structure, location and intensity errors of identified features (Jolliffe and Stephenson, 2012).

7.3. *Understanding the development mechanisms of PLs*

Despite the large number of case studies conducted, a comprehensive understanding of the mechanisms involved in the formation and intensification of PLs is yet to be obtained. The analysis of the energy budget of PLs is a valuable approach to acquire knowledge about the energy sources, sinks and conversions involved in PL development. Although some work has already been done in that direction, there is still room for improvement. The energy budget equations should ideally be applied to the output from high-resolution coupled ocean-atmosphere models as they represent the energy fluxes between the ocean and

the atmosphere, thus capturing the atmosphere-ocean feedbacks involved in PL intensification. The energy budget of a PL could be analysed using local energy cycle equations such as those developed by Nikiéma and Laprise (2013), which have been successfully applied by Clément et al. (2017) to analyse the role of different mechanisms in the development of an extratropical cyclone.

7.4. Analysing the impact of climate change on the climatology of PLs

Finally, the impact of climate change on the spatio-temporal distribution of PLs is an area of research that has received attention in recent times. With the advent of high-resolution RCMs, it has become possible to analyse the future climatology of PLs. The use of coupled atmosphere-ocean RCMs to analyse the impact of climate change on PLs is a promising area of research, although it has been limited so far by the high computational cost. Global climate models (GCMs), which have long been excluded from the study of PLs due to their mesh coarseness, could prove to be a valuable tool to analyse the climatology of PLs. In effect, with the development of variable-resolution GCMs, it may be computationally feasible to study the climatology of PLs in a particular region using such GCMs. These advances in climate modelling will likely lead to prolific research on the impacts of recent and future climate change on the climatology of PLs, and its potential characteristics in terms of occurrence, intensity, lifetime, track and frequency.

Note

1. The Nordic Seas include the Norwegian Sea, the Greenland Sea, the Iceland Sea and the Barents Sea.

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[EO:EUM:DAT:METOP:OSI-104](https://navigator.eumetsat.int/product/EO:EUM:DAT:METOP:OSI-104)), and the multi-radar mosaic has been downloaded using MET Norway Thredds Service (<https://thredds.met.no/thredds/catalog/remotesensing/reflectivity-nordic/catalog.html>). ERA5 reanalysis from ECMWF is available at <https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5>. The authors are also grateful for the Global Self-consistent, Hierarchical, High-resolution Geography Database (GSHHG), developed and maintained by Dr. Paul Wessel, SOEST, University of Hawai'i, and Dr. Walter H. F. Smith, NOAA Laboratory for Satellite Altimetry. The GSHHG is available at <https://www.ngdc.noaa.gov/mgg/shorelines/>. The GSHHG has been used to represent the coastlines of the Northern and Southern Hemispheres. The authors would like to also thank Georges Huard and Frédérik Toupin for maintaining an efficient and user-friendly local computing facility, and Katja Winger for downloading and preparing available ERA5 reanalyses at the ESCER centre.

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